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THE GEOLOGY OF CENTRAL MINAS GERAES, BRAZIL

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PART II

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PRINCIPAL MINERAL DEPOSITS

IRON ORES

On the basis of their physical properties, which to some degree separates them also according to chemical composition, the iron ores[1] of Minas Geraes may be classified as follows.

[1] H. K. Scott, "The Iron Ores of Brazil," *Jour. Iron and Steel Institute*, LXI, (1902), 248; O. A. Derby, "The Iron Ores of Brazil," *Iron Ore Resources of the World*, p. 313, Stockholm, 1910; C. K. Leith and E. C. Harder, "The Hematite Ores of Brazil and a Comparison with the Hematite Ores of Lake Superior," *Econ. Geol.*, VI, (1911) 670–86; E. C. Harder, "The 'Itabirite' Iron Ores of Brazil," *Econ. Geol.*, IX, (1914) 101–11; E. C. Harder, "The Iron Industry in Brazil," *Bull. Am. Inst. Mg. Engrs.*, October, 1914.

Original or Bedded Ores
- Hard massive ore
- Soft powdery ore
- Laminated or thin-bedded ore

Concentration Ores
- Canga
- Stream sand and gravel ores
- Rubble ore
- Enriched itabirite
- Leached carbonate

Of these the bedded ores, especially the hard massive ores, are by far the most important. Of the concentration ores the canga is the principal type.

Bedded ores.—Bedded iron ores are found in many places in central Minas Geraes, but the most important deposits occur in the region surrounding the headwaters of Rio Piracicaba, Rio Carmo, Rio das Velhas, Rio Santo Antonio, and Rio Paraopeba, together with their various branches and tributaries. This is the region in which the Itabira iron formation has survived the extensive denudation which has swept away the post-Archean formations from much of eastern Brazil, and where it still remains as a thick formation over considerable areas. The ore bodies occur at various points along the belts of iron formation already described, both in the Itabira iron formation proper, and in the iron-formation lenses in the Piracicaba schist. They are not irregularly scattered throughout the iron-formation areas, but are rather grouped in certain portions with barren areas between. Thus groups of iron-ore deposits occur north of Ouro Preto on the east slope of the Serra do Caraça;[1] on the Rio Piracicaba west and northwest of the town of Villa Rio Piracicaba; on the prominent peaks at Itabira de Matto Dentro; along the ridge running from Sabara northeastward to Caethé and southwestward beyond Bello Horizonte; at various points in the area west of Itabira do Campo, and in the region west of Burnier and north of Gongonhas do Campo. Smaller, less important groups occur elsewhere in the district.

The bedded ores occur as beds or lenses in the iron formation. Some are thin and continuous for long distances, while others are

[1] See map on p. 347, Part I of this article.

thick and of small diameter. Of the three classes of bedded ores, hard massive ore and the peculiar soft powdery ore are the most widespread, the latter probably being the more common of the two. Laminated ore is less widespread, but generally occurs in larger deposits.

The hard ore of central Minas is of several varieties, the most common of which is a slightly specular blue hematite resembling

FIG. 12.—The Peak of Caué rising above the town of Itabira de Matto Dentro. A large body of hard ore.

some of the specular hematite ores on the Marquette range in Michigan. Such ores occur in many places in the district. Locally there are deposits consisting of coarsely specular hematite with scattered crystals of martite, and with these are frequently associated ores composed largely of coarse granular masses of martite, or magnetite not yet completely altered to hematite. Such phases are due probably to greater metamorphism and are most common in the eastern part of the district. In the central and western part of the district there occur a few deposits of a tough, fine-grained, almost amorphous hematite, or hematite intermixed with some limonite.

Some hard ore deposits, or parts of these deposits, are so massive that the bedding is barely distinguishable while other portions may be distinctly bedded. The material is generally well consolidated, and some phases, especially the amorphous variety, are of extraordinary hardness and toughness.

Of the different varieties of hard ore the finely specular type is the highest grade. It generally averages between 69 and 70 per cent in metallic iron and rarely has more than 0.025 per cent phosphorus and frequently runs as low as 0.003 per cent phosphorus. Other impurities are practically absent.

The soft, powdery ore deposits consist largely of unconsolidated material of great fineness, most of it so fine-grained that it will easily pass through a 100-mesh screen. Although fine, the material is almost entirely crystalline, quite different from the amorphous powder which occurs with much of the soft ore on the Mesabi Range in Minnesota. Some of the powdery ore is slightly consolidated when moist, but readily crumbles when dried. Most of it, however, is soft in the moist state like fine moist sand.

The powdery ore is dark blue in color like the hard ore and resembles it in composition, though in general perhaps of slightly lower grade as it is more frequently intermixed with a sprinkling of quartz sand. It averages between 67 and 69.5 per cent metallic iron and contains up to 0.05 per cent of phosphorus, though it may run as low as 0.004 per cent of phosphorus. When the amount of quartz sand increases in the soft ore the metallic iron content decreases until the arbitrary figure of 50 per cent metallic iron is reached, when the material is classified as itabirite. Some of the soft ores, just as in the case of some of the hard ores, are almost without evidence of bedding, while others are distinctly bedded and even finely laminated. Where quartz sand is intermixed with the soft ore the bedding is well marked.

Hard ore and soft ore are generally more or less associated with each other. In general it may be said that massive specular ore and soft, powdery ore are hard and soft phases of the same material (Fig. 13). A deposit composed principally of hard ore may have lenses of soft ore or irregular masses of soft ore scattered through it, or portions of it may be composed of a mixture of hard

ore fragments and soft ore, the latter acting as a matrix. On the other hand, deposits consisting largely of soft ore may have hard ore lenses interbedded with the soft ore.

The soft, powdery ore is known among the natives of the iron region as "jacutinga," and since this term has also crept into the literature its meaning should be clearly defined and its misuse avoided. The term "jacutinga" was applied by Hussak[1] to

FIG. 13.—The Peak of São Luis near the village of Agua Quente. In front of the peak is a plain surfaced by canga. The steeply inclined beds of iron formation which appear in the peak continue under the canga covering of the plain. In the peak they consist partly of itabirite and partly of hard ore; under the plain the same beds consist of soft, powdery ore and soft itabirite. On the extreme left is the Caraça quartzite range.

certain gold-bearing portions of the iron formation. The jacutinga thus designated occurs in the form of shoots and irregular masses inclosed in itabirite from which it differs in a minor way only. It has a variable composition, and its principal characteristic is a fine

[1] E. Hussak, "O Palladio e a Platina no Brazil," *Annaes da Escola de Minas de Ouro Preto*, N. 8 (1906), p. 96, tr. by Miguel A. R. Lisboa, and Manoel A. R. Lisboa.

micaceous texture. It consists mainly of soft hematite in which quartz occurs in varying abundance together with talc, kaolin, mica, and earthy pyrolusite and frequently tourmaline. Presumably the term as used by Hussak includes those portions of the iron formation which have been affected by gold-bearing solutions, and it seems best to limit its use in the future to this phase and to avoid its use in the iron-ore terminology.

The bedded ores of the third class, the laminated, are distinct from those of the other classes and more closely resemble the itabirite in texture. These ores might be termed itabirite with little or no quartz sand. They occur in large lens-like deposits in the iron formation, as do the hard ores and soft ores, but the boundary between one of these lenses and the itabirite of the iron formation which incloses it is less well defined and more irregular. While the hard ore and soft ore deposits generally have a sharp contact with the inclosing itabirite, the laminated ore deposits generally have a zone of gradation with the itabirite and also have interbedded masses of itabirite within them. The separation of itabirite from laminated ore is frequently quite arbitrary, being controlled by the percentage of metallic iron.

Lenses of hard ore or soft powdery ore occur interbedded within laminated ore deposits in the same way that they occur in the itabirite portions of the iron formation. In such cases the contact with the hard ore is well defined, but that with the soft ore less so.

Laminated ore is a very thinly bedded, porous ore which is quite friable, breaking into thin plates along bedding planes, owing to the fact that the porosity is concentrated along these planes. The ore is red or blue hematite, much of it in the amorphous form, with a considerable percentage of limonite. The different varieties of laminated ore are generally found interbedded or interlaminated. Certain beds may be red, others blue, and still others yellow, the differences being the result of hydration and slight admixtures of clay and other impurities. Frequently siliceous beds of different thicknesses will be found interbedded with pure ore.

The laminated ore is not so high in grade as the hard ore and soft ore, for the metallic iron content of its richer portions varies

between 63 and 67 per cent, while the hydrated portions frequently contain as high as 0.3 per cent phosphorus and rarely less than 0.05 per cent phosphorus. It is mainly a non-Bessemer ore. With increasing quartz sand the content of metallic iron decreases until the material becomes itabirite. Non-hydrated, laminated ore resembles soft, powdery ore in composition.

The phosphorus content of the surface ores is always higher than that of the ore underground. Where surface ore contains 0.12 to 0.18 per cent phosphorus, underground ore may contain only 0.04 to 0.08 per cent. This is due to the concentration of phosphorus during the process of weathering. This same process also concentrates the iron in the uppermost foot or two, hardening it and decreasing the pore space.

The size of the different ore deposits varies from mere seams inclosed in itabirite to beds several kilometers in length and several hundred meters in thickness. The largest known deposits are of laminated ore, but there are also hard ore deposits of large size. The largest known laminated ore deposit in the district is that of Alegria, where continuous outcrops occur over a distance of more than 4 kilometers along the strike of the beds and for fully 1,700 meters across the beds. With an average dip of 35° the thickness would be at least 1,200 meters.[1] This great thickness, however, is not entirely of pure ore, but will possibly be found to include numerous minor beds of itabirite now concealed at the surface. The largest body of hard ore in the district, which, however, has a considerable admixture of soft ore and of hard ore fragments intermixed with soft ore, is that of Periquito near Itabira de Matto Dentro. It has a length of 1 kilometer and a maximum thickness at right angles to the beds of 250 meters. Other large deposits of hard ore are known in which the proportion of soft ore is very small. Such are the peak of Caué (Fig. 12), the deposits on Serra de Conceiçao and Serra do Esmeril near Itabira de Matto Dentro, Itabira Peak near Itabira do Campo, and Morro Agudo Peak near Villa Rio Piracicaba. Deposits consisting largely of pure soft powdery ore are as a rule much smaller than hard ore deposits but more numerous.

[1] See Fig. 4 on p. 354, Part I of this article.

Concentration ores.—Concentration ores occur in the neighborhood of other ore deposits or of iron-formation beds by the disintegration of which they are formed. They are in the form of surface deposits and do not go to a great depth. Of such deposits the canga ores are the most widespread, occurring as surface blankets of great horizontal extent over a large portion of the areas of iron formation, as well as over areas of other rocks bordering the iron-formation

FIG. 14.—The Fazenda de Alegria. The prominent mountain is the Serra do Caraça. The lower range of hills in the middle distance are the hills of laminated iron ore. The ranch house is built upon canga-covered Piracicaba schist.

belts. They are composed of material resulting from the disintegration, not only of the iron ore portions of the iron formation, but from the itabirite phase as well.

Canga consists of different sized fragments of iron ore mixed in places with fragments of itabirite, the whole cemented together by iron oxide. The incorporated fragments vary in size from sand grains to bowlders weighing perhaps several tons. Some of them consist of pure massive ore, others of laminated ore or itabirite. The fragments have been mechanically concentrated and have

been cemented together by iron oxide deposited from solution, both fragments and cementing material being derived from the same source—the iron formation. The cementing material is largely hydrated hematite and limonite, red or yellow in color, giving different portions of canga deposits colors varying from light yellow to dark red. The fragments decrease in size and abundance as the distance from the source increases, so that in one place a canga deposit may consist largely of cemented fragments and elsewhere of finely textured chemically deposited iron oxide. The latter is most common over areas outside the iron formation or over soft portions of the iron-formation belts where erosion yields no coarse fragments. In the latter localities, however, much fine fragmental material is intermixed. In places at a distance from the iron-formation areas the canga is very low grade, containing much clay and quartz sand, the iron oxide present being principally limonitic cement.

The ordinary canga of which the great portion of the canga deposits are composed averages between 60 and 65 per cent metallic iron and up to 0.3 per cent phosphorus, rarely containing less than 0.1 per cent phosphorus. This concentration of phosphorus is due to the same cause as the concentration of the phosphorus in the surface laminated ore. The outlying portions of canga deposits at a distance from the iron-formation belt are so low in metallic iron that they do not constitute an iron ore.

A canga blanket may vary in thickness in different portions from a few centimeters to 20 meters or more. It is generally thickest on the lower slopes and at the base of iron-formation ridges, where the conditions for accumulation are best, and decreases in thickness both up the slope and toward the valley. A single continuous blanket of canga may cover several square kilometers.

Stream deposits of iron oxide in the form of sand, pebbles, and larger fragments occur in greater or less purity along most of the streams rising in or flowing through areas of iron formation. The material is distributed over the width of the valley floors and in places occurs in terraces several meters above the present valley bottom. The stream deposits do not constitute a very high-grade ore, always being mixed with more or less quartz sand and other

foreign material. They are of little or no importance as an ore reserve.

Rubble ore deposits occur on the hill slopes below outcrops of hard ore. They are simply talus accumulations consisting of high-grade iron ore, and frequently are large enough to be of some importance for mining purposes. The fragments composing them may vary in size from pebbles to fragments of several tons. In

FIG. 15.—A cliff of canga on the side of the "Grotto" in Esmeril near Itabira de Matto Dentro. The canga blanket here reaches a thickness of about 15 meters.

many places, however, one finds rubble ore deposits with fragments remarkably uniform in size. Rubble ores have practically the same composition as the hard ore deposits with which they are associated.

Small deposits of enriched itabirite occur locally, but are of little importance. They might be termed canga deposits formed *in situ*, as they occur at the surface in itabirite areas where the leaching of silica has been sufficient to leave the residual material concentrated as iron ore. Frequently the iron oxide itself has been dissolved and redeposited in the form of limonite in pores and along

cracks as veins and stringers. In some places manganese oxide (psilomelane) is associated with this redeposited material.

The lenses and beds of carbonate rocks in the lower part of the Piracicaba formation in places contain a noticeable percentage of iron and manganese carbonates. On decomposition these form impure residual deposits of iron and manganese oxides.

Origin of the bedded ores.—From the structure and shape of the bedded ore deposits and their relation to the inclosing rocks it is necessary to conclude that they are original sedimentary deposits, laid down in a measure similar to limestone beds. They are simply a part of the iron-formation beds, similar in origin to the itabirite with which they are associated, except that they contain a smaller proportion of mechanical sediments, such as quartz sand. It is, therefore, best to explain the origin of the iron formation as a whole and note the differences in origin between the itabirite portions and the iron ore portions.

As has been stated, the iron formation is, in the main, an iron-oxide-bearing sandstone or quartzite. It varies in different portions from a rock consisting largely of quartz sand to one consisting of pure iron oxide. The siliceous portions of the iron formation are called itabirite, just as the siliceous portions of the Lake Superior iron formation are called ferruginous chert, jaspilite, or taconite. The portions rich in iron oxide are termed massive ore, powdery ore, or laminated ore, according to their texture. The line between the itabirite and the iron ore, especially the laminated ore, is arbitrary, depending on the percentage of metallic iron present. At the present state of the iron industry, iron formation with 50 per cent or more of metallic iron may be termed iron ore and iron formation containing below 50 per cent metallic iron itabirite. In the Brazilian iron formation there are no equivalents of the cherty iron carbonate and the greenalite rock of the Lake Superior district which in that region represent the original rock from which the iron ore, ferruginous chert, and associated rocks have been formed by weathering. In the Brazilian iron fields there has been no such alteration, the itabirite and iron ore being the original rocks.[1]

[1] C. K. Leith, and E. C. Harder, "The Hematite Ores of Brazil and a Comparison with the Hematite Ores of Lake Superior," *Econ. Geol.*, VI (1911) 670–86; E. C. Harder, "The 'Itabirite' Iron Ores of Brazil," *Econ. Geol.*, IX (1914), 101–11.

The iron formation has all the characteristics of a sedimentary deposit, although, just as in the rest of the Minas series, fossils have not been recognized in it. Portions are very massive and exhibit little or no evidence of bedding, as is frequently seen in massive quartzites, while other portions are thin-bedded or finely laminated, as shales and fine-grained sandstones. In the thinner-bedded portion the bedding or lamination planes are for the most

FIG. 16.—The Fazenda de Alegria. The highest peaks behind are of Caraça quartzite. The strong range of hills in front are of iron formation. To the left of the deep notch they are chiefly of itabirite; to the right of it dominantly of laminated iron ore.

part sharply defined. The stratification in places is rendered very conspicuous for the reason that in the itabirite portions of the formation the iron-oxide layers are so frequently separated by thin partings of white quartz sand. These sandy partings, as planes of weakness, control to a considerable extent the weathering and fracturing of the beds. While the main bulk of the iron formation consists of sandy iron-oxide-bearing beds, there frequently occur, interbedded with these, shaly or schistose iron-oxide-bearing beds

showing that, while the mechanical sediments consisted mainly of quartz sand, clay also was frequently deposited in certain localities.

Thus it is clear that in structure and general occurrence the iron formation is so like ordinary sediments that there is little doubt as to its sedimentary origin. It remains, however, to account for its peculiar and unique composition and also for its enormous mass. It is difficult to picture a source for such an enormous amount of iron oxide as is contained in the iron formation, and even if an adequate source of supply be found, it is not so easy to see how the iron oxide came to be deposited locally in masses hundreds of meters in thickness with less than $\frac{1}{2}$ of 1 per cent of siliceous sediments intermixed.

It is apparent from the composition of the Caraça quartzite that during the first stages of the sedimentary deposition the processes of decomposition greatly exceeded in effectiveness the processes of mechanical disruption, for, instead of arkose sediments, the Caraça quartzite is made up of the products of mature weathering. This may be ascribed to a long period of decomposition without transportation, resulting in the accumulation of a thick mantle of residual material on the land surface, or it may be ascribed to a warm, moist climate, or to extensive wave action with practically continuous sedimentation. The material of which the sediments are composed was derived from the decomposition of rocks of the basement complex, such as granite, gneiss, amphibolite, and crystalline schists, with some basic intrusives. The principal minerals associated in the make-up of these rocks are quartz, feldspar, biotite, hornblende, and some pyroxene, whose end products after weathering are quartz, kaolin, hydrated iron oxide, and some aluminum hydroxide. These four constituents, therefore, probably made up the bulk of the decomposition products.

During the sedimentation which followed the post-Archean erosion period the first material to be deposited in central Minas Geraes was quartz sand, from which it is natural to suppose that the finer products of the decay of the basement complex—the kaolin and hydrated iron oxide—were carried farther out to sea, while the coarser quartz sand was deposited nearer the shore. This process must have been long continued, for the Caraça

quartzite is thick. Following the Caraça quartzite is the Batatal schist, the finer material of which implies a slackening in the vigor of sedimentation, whether because the land from which the sediments were being derived was becoming lower, or because the source of the sediment was becoming more remote owing to an advancing shore line, or because of deepening of waters or shifting of currents. The Batatal schist is thin, indicating that this stage was not long enduring. It was followed by local flows of basic lava. And then came the deposition of the iron formation. The hypotheses to explain this remarkable formation naturally are of two sorts: (1) that the iron oxide was a mechanical sediment washed in like the mud of the schist and the sand of the quartzite; (2) that it was precipitated from solution through either chemical or biochemical agencies.

Following the hypothesis that the iron oxide was a mechanical sediment, the source for the iron oxide should naturally be the residual material resulting from the thorough weathering of the rocks of the basement complex. Besides this hydrated iron oxide, the residual material should contain kaolin and quartz in somewhat greater proportion than the iron oxide. Under any hypothesis the sediments which resulted in the Caraça quartzite and the Batatal schist came from the sorting of thoroughly weathered material by stream and wave action. The Caraça quartzite represents a time when only material of the coarser sort, the sand grains, could find rest in the area under consideration. The finer material was swept farther out to sea to find lodgment in deeper and quieter waters. With the lessening of the vigor of the agents of sedimentation came the thin Batatal schist. At this time fine mud was being washed into the sea and deposited, while the coarser sand was either not washed into the sea or was left still nearer the shore which may have changed its position in the meantime. With the kaolin much iron oxide must also have been washed into the sea and not a little of it deposited with the mud.

Thus far the process has been the ordinary one of decay of igneous rocks and the deposition of much of the resulting material as clastic sediment. The unusual feature to be explained is what very peculiar condition obtained to cause the deposition of ferric

hydroxide pretty generally over the region and in some places the building up of a series of beds of ferric hydroxide in a high state of purity which sometimes totals several hundred meters in thickness. The completeness of the separation, as evidenced by the fact that much of the ore now contains 69 per cent iron out of a possible 70 per cent for chemically pure ferric oxide, and the vast extent of the iron formation are very hard to understand. The iron oxide resulting from the decomposition of granite, gneiss, amphibolite, and crystalline schists should be in a more or less finely divided condition, and as such should not be very generally separated from the sand and clay by any sorting action of running water or ocean waves, except in so far as affected by a higher specific gravity which might result in small local accumulations of iron sands, or in the association of smaller particles of iron oxide with larger particles of other materials in the sediments. The iron oxide should be with the schists and quartzites. Great masses of iron oxide containing less than $\frac{1}{2}$ of 1 per cent of impurity do not seem possible as the result of mechanical separation, though it is not impossible that a small percentage of the iron oxide in the iron formation may have been derived thus.

The alternative hypothesis is that the iron oxide was precipitated from solution following the formation of limestones as an analogy. This escapes the serious difficulties of separation of materials. As in the case of the limestones, to account for the purity it is only necessary to suppose that the deposition of the iron compound occurred in clear seas where comparatively little clastic sediment was being brought in. While the thickness of iron formation is great, it is exceeded by many limestones.

It is the nature of the precipitation that leaves the widest field for speculation. There are two possibilities: (1) the iron may have been precipitated directly from solution by purely chemical means, or (2) it may have been abstracted from the water by organisms and deposited as a sediment. In either case the iron may be supposed to have been in solution chiefly in the form of ferrous carbonate and to have been thrown down as ferric hydroxide.

There are various chemical reactions which can cause the precipitation of ferric hydroxide from solutions of ferrous carbonate,

of which in nature oxidation of the carbonate with the resulting hydrolysis is perhaps the most likely possibility. But such a reaction should necessitate a notable quantity of iron compound in solution. Van Hise and Leith object that river and sea-waters do not contain the requisite amount of iron.[1] Their suggestion that the iron in the various iron formations of the Lake Superior region has come from associated basaltic lavas, either from the magmatic waters or from chemical reactions between the hot basic lavas and the sea-water,[2] hardly seems applicable to the Brazilian iron formation, since nothing in the nature of basic lavas has been found within the sedimentary series of the iron-ore district with the exception of the serpentinized remains of one small flow near Cattas Altas. This is very insignificant in extent in comparison with the iron formation, and furthermore extensive lenses of iron formation occur higher up within the Piracicaba schist with which there are no igneous rocks associated. In chemical precipitation there is likewise to be considered the fact that chemical reactions would be likely to produce other precipitates in addition to ferric hydroxide, such as aluminum hydroxide and calcium compounds, which would result in the formation of impure deposits of iron ore rather than in thick beds and lenses of pure ferric hydroxide.

The other possibility is that the ferric hydroxide was thrown down by organic action. It is now known that much of the bog iron ore being formed in lagoons at the present time is the result of the activity of a certain group of bacteria known as the iron bacteria. The iron bacteria include many individual species, of which the thread bacteria *Chlamydothrix*, *Gallionella*, *Spirophyllum*, *Crenothrix*, and *Clonothrix*, and the coccus form *Siderocapsa* have perhaps been most carefully studied.[3] While the different species have individual morphological peculiarities of their own, the type

[1] C. R. Van Hise and C. K. Leith, "The Geology of the Lake Superior Region," *Mon. 52, U.S. Geol. Surv.* (1911), pp. 503–6.

[2] *Op. cit.*, pp. 506–18.

[3] Hans Molisch, *Die Eisenbakterien*, p. 10, Jena, 1910; D. Ellis, "A Contribution to Our Knowledge of the Thread Bacteria," *Centralbl. für Bakt.*, Abt. II, Bd. 19 (1907), p. 502; Abt. II, Bd. 26 (1910), p. 321.

as a whole possesses certain general physiological characteristics. They all live in clear water, either standing or running water. Lieske states that he has never found any in turbid water, nor in waters containing a great deal of organic matter.[1] They live in waters containing iron compounds in solution which it is claimed by Winogradsky they utilize according to the following reaction: $2FeCO_3+O+3H_2O=2Fe(OH)_3+2CO_2$. Heat is liberated by this reaction, and this energy together with the carbon dioxide developed is utilized by the bacteria to sustain life.[2] Ferric hydroxide is left behind and may accumulate. Other investigators, like Molisch,[3] claim that ferrous compounds are not necessary for the physiological processes of these organisms and that organic compounds other than carbon dioxide must be present for their use. Nearly all agree, however, that their activity results in the accumulation of deposits of ferric hydroxide in many places. As the result of this activity the water pipes of cities where the water contains a considerable amount of ferrous carbonate have sometimes been completely closed.[4] That certain limonite deposits have been produced in this way is evidenced by the fact that in them large numbers of the sheaths of these bacteria have been found.[5] To quote from Lafar:

> The decomposing power of these organisms is very great, the amount of ferrous oxide oxidized by their cells being a high multiple of their own weight. This high chemical energy on the one hand, and the inexacting demands in the shape of food on the other, secure to these bacteria an important part in the economy of nature, the enormous deposits of ferruginous ocher and bog iron ore, and probably certain manganese ores as well, being the result of the activity of the iron bacteria.[6]

[1] Rudolf Lieske, "Beiträge zur Kenntnis der Physiologie von Spirophyllum ferrugineum (Ellis), einem typischen Eisenbakterium," *Jahrb. für wissenschaftliche Botanik*, XLIX (1911), 91–127.

[2] S. Winogradsky, "Über Eisenbakterien," *Botan. Zeitung*, Bd. 46 (1888), p. 261.

[3] Hans Molisch, *op. cit.*, p. 44.

[4] F. Lafar, *Technical Mycology*, I (1898), 361; also I (1910), 272.

[5] A. Fischer, *The Structure and Functions of Bacteria*, p. 69, tr. by A. Coppen Jones, Clarendon Press, Oxford, 1900.

[6] F. Lafar, cited by Van Hise and Leith, *op. cit.*, *Mon. 52, U.S. Geol. Surv.* (1911) p. 519.

Such a process as this would seem to offer a possible clue to the origin of the Itabira iron formation. As these bacteria thrive best in clear waters, the low proportion of clastic sediment in the formation is natural enough. If the deposition in general took place in lagoons and embayments and the thicker and more siliceous portions of the formation developed near the river mouths, in regions of currents, or other favorable localities, and the thinner, more uniform portions of the formation represent deposits formed

FIG. 17.—A view of the Peak of Conceição near Itabira de Matto Dentro, from the summit of Caué. The iron formation forms a continuous belt between Caué and Conceição.

farther off shore or in quiet waters near shore, the great and sudden variations in the thickness of the iron formation at various points may perhaps be accounted for. Much of the ferric hydroxide may have been formed as a flocculent precipitate in the sluggish river waters carrying little or no clastic sediment, and later deposited in a thick series of delta beds at the debouchures of the streams. Such may be the great accumulations of iron formation at Alegria, Gandarella, and elsewhere. Whatever the nature of this sedimenta-

tion, the field evidence appears to indicate that the deposition of the iron-oxide beds in these localities of unusual thickness took place with comparative rapidity.

While certain thin discontinuous layers of iron formation might be attributed to chemical percipitation, it is difficult to realize how great thicknesses over large areas could have been thus formed. Little is known concerning the concentration of iron compounds in solution necessary for chemical precipitation. It probably varies greatly with varying conditions, and the special conditions which would have to be assumed as causing the deposition of the Brazilian iron formation must have extended over large, as well as widely scattered, areas since this formation is not only found throughout a considerable area in Minas Geraes but also exists in the extreme western part of Brazil, nearly a thousand miles to the west. On the other hand, there is experimental evidence that bacteria do precipitate iron oxide out of very dilute solutions, and it is only necessary to assume the presence of large numbers of these micro-organisms in scattered localities to account for the presence of the iron formation. Unfortunately the metamorphism which the iron formation has suffered makes it impossible to recognize organic remains, if such were originally present. Recognizable bacterial remains, however, are stated to be extremely short-lived and it is difficult to identify them even in modern bog-iron-ore deposits. There is also to be considered the fact that nowhere in the ocean do we know of extensive deposits of ferric hydroxide being formed at the present time. This objection, however, is equally valid in case of ferric hydroxide precipitated either chemically or biochemically, and therefore must be considered under either hypothesis. In the formation of bog ores on continental areas bacteria are known to play an important part, but chemical precipitation is probably effective also, and it is difficult to say which plays the principal rôle. In spite of various uncertain factors, however, the deposition by micro-organisms seems more easily to explain the various unusual phases of the Brazilian iron-formation sedimentation, and we prefer to adopt this hypothesis, at the same time, however, realizing fully the inadequacy of our present knowledge concerning bacteria as a

geological agency, and the necessity for further investigation on this most important subject.

MANGANESE ORES

With the exception of gold, manganese ore is the most important of the metalliferous mineral products of Brazil at the present time. Two mines in central Minas Geraes, the Morro da Mina mine north of Lafayette, and the Wigg mine east of Miguel Burnier, are in continuous operation, while several smaller mines produce ore intermittently. Among these are the Rodeio mine near Kilometer 508, east of Miguel Burnier on the Ouro Preto branch of the Central Railroad of Brazil, the Cocuruto mine southwest of Lafayette, and the Queluz das Minas mine near the Morro da Mina mine north of Lafayette. Many abandoned and inactive mines occur in the general vicinity of the mines mentioned above.

The manganese deposits of central Minas Geraes may be separated into two distinct classes: (1) those occurring in the basement complex and (2) those occurring in the overlying sediments. The deposits found in the region around Lafayette belong to the first class, while those occurring along the Ouro Preto branch railway east of Miguel Burnier are of the second class. The centers of these two districts, having distinct types of deposits, are not more than 25 kilometers apart, and some of the mines in the one can be seen from the other. The deposits have been described in detail by Derby[1] and by Scott,[2] so that only a general outline need be given here in order to show their relation to the general stratigraphy.

The manganese deposits which are found in the basement complex consist of large irregular masses of manganese oxide inclosed in, or bounded by, gneiss, granite, or crystalline schist. Individual masses such as that at Morro da Mina may be more than a hundred meters in their longer diameter. While irregular in shape, they are usually somewhat elongated, suggesting lenses. They occur

[1] O. A. Derby, "On the Manganese Ore Deposits of the Queluz (Lafayette) District, Minas Geraes," *Am. Jour. Sci.*, XII (1901), 18–32; "On the Original Type of Manganese Ore Deposits of the Queluz District, Brazil," *ibid.*, XXV (1908), 213–16.

[2] H. K. Scott, "The Manganese Ores of Brazil," *Jour. Iron and Steel Inst.* (1900), p. 179.

scattered through the basement complex without any apparent regularity, but most of them appear to have either gneiss or crystalline schist on one or both bounding walls.

The manganese oxide composing these lenses is usually in the amorphous form, occurring mainly as psilomelane and wad, though pyrolusite also is found with these. According to detailed studies made by Dr. Derby,[1] it appears that these oxides are surface decomposition products of other manganese minerals which have in one or two cases been encountered below the zone of oxidation. Of these minerals the principal ones are the manganese silicates, tephroite and spessartite, and with these occur rhodochrosite, the manganese carbonate, and sparingly rhodonite, another manganese silicate. These minerals occur intimately intermixed in varying proportions, one being more abundant in one place and another elsewhere, and together they form a reddish manganese silicate and carbonate rock. The relation of the manganese rock to the inclosing crystalline rocks has not been definitely determined; it may be interlayered with the gneiss or crystalline schist or perhaps it is intrusive into them.

From the one or two instances noted it is judged that all the manganese oxide deposits in the areas of basement complex are surface oxidation products of such masses of manganese silicate and carbonate rock. In many of the deposits where the original rock has not been encountered, the oxide ores have textures which are duplicated in the manganese silicate and carbonate rock elsewhere, and therefore suggest a similar origin. During the process of decomposition more or less solution and redeposition takes place, with the result that certain portions of a deposit are composed of high-grade manganese oxide, while other portions contain admixtures of other products of decomposition, such as clay and quartz sand. Most of the ore is hard, but soft material, mainly wad and pyrolusite, also occurs abundantly, being irregularly intermixed with it.

Manganese ores associated with igneous rocks, such as those described above, occur abundantly in India and are also found locally in the eastern United States.

[1] O. A. Derby, *Am. Jour. Sci.*, XXV (1908), 215–16.

The manganese deposits in the sedimentary series occur as definite beds associated with the iron formation. The principal bed, that on which the Wigg mine is situated, is 3 or 4 kilometers in length, and at its maximum reaches a thickness of over 2 meters. It strikes in an east-west direction parallel to the strike of the inclosing sediments and corresponds with them in dip, making it apparent that the manganese ore bed was laid down as a sedimentary bed just as the inclosing rock. The bed at the Wigg mine is bounded on one side by soft itabirite, with a contact zone of mixed soft hematite and manganese oxide, and on the other side by a ferruginous schist associated with the iron formation.

The manganese bed at the Rodeio mine is of smaller horizontal extent but of greater thickness than that at the Wigg mine and shows less definitely its relation to the inclosing rocks. In the case of both of these deposits, beds of carbonate rocks consisting of a mixture of calcium, magnesium, iron, and manganese carbonates are found in the vicinity. These, however, occur at different horizons from the manganese beds.

The manganese ores associated with the sedimentary rocks consist of finely crystalline or amorphous manganese oxides, probably largely a mixture of pyrolusite and psilomelane. From their occurrence it must be assumed that they are similar in origin to the associated rocks, that is, that they are original sedimentary deposits of manganese oxide which have been somewhat altered and recrystallized by subsequent metamorphism. The source of the manganese is doubtful, but it may very well have been derived from deposits of manganese ore in the basement complex such as now occur to the south near Lafayette. Decomposition of such deposits may have yielded a large amount of residual manganese oxide which was worked over, transported, and deposited as beds or lenses in the sedimentary series. Their origin would thus be very similar to that of the iron ores of central Minas Geraes with which they are closely associated.

Sedimentary manganese ore beds similar to those of central Minas Geraes are found in many places on other continents, probably the best known of them being those in northern Arkansas, those in central Chile, those in the Caucasus near the Russian boundary, and those in western Arabia.

GOLD

Gold was known to occur in Brazil even in the early days of its settlement and it was the search for this metal which brought many of the explorers and settlers to the new country. However, not until the beginning of the eighteenth century was the great gold field of Brazil, that occurring in central Minas Geraes, actively worked. It was at this time that the towns of Sabara, Marianna, and Ouro Preto, the latter then known as Villa Rica, were founded

FIG. 18.—Ouro Preto, former capital of Minas Geraes. A city built by the gold-mining industry. The hills on all sides show the results of placer mining.

in rapid succession. A period of great excitement and activity followed, which continued in waves as districts were extended and new ones discovered, almost to the middle of the nineteenth century. From that time on the industry rapidly decreased in importance until 1889 when gold mining as a general industry practically ceased as a result of the abolition of slavery. At the present time there are only two important gold mines operated in Minas Geraes. These are the Morro Velho mine at Villa Nova de Lima near Bello Horizonte and the Passagem mine at Passagem near Ouro Preto.

Nearly all the early workings were shallow open cuts or shafts, or short tunnels. Later development led to the operation of a few deep mines, such as the Descoberto mine near Sabara, the Gongo Socco and São Bento mines between Caethé and Santa Barbara, the Pary mine near São Francisco, and the Santa Anna and Maquiné mines near Marianna. The shallow workings cover large areas and are widely distributed in central Minas, groups of them being found in the vicinity of nearly all the small towns and villages, for in most cases it was the gold mines that caused the founding of the villages. Some of the areas of old abandoned workings cover many hundreds of acres.

Gold occurs in Minas Geraes in three different associations: (1) in quartz or sulphide veins, (2) disseminated in the iron formation and in the canga derived from it, and (3) in stream gravels.

Gold-bearing quartz veins occur with different relations; some of them are strike veins, others occur along bedding or schistosity planes; some are long and continuous, others short and lens-like. In some places groups of short parallel or intersecting veins are found, while elsewhere single isolated veins of considerable extent occur.

Quartz or sulphide veins containing gold may occur in the basement complex or in any of the sedimentary formations. The Morro Velho mine at Villa Nova de Lima is operating a large vein in the Piracicaba schist on which they have descended for a vertical distance of more than 1,700 meters, deeper than any other gold mine in the world. The Passagem mine near Ouro Preto is working on an irregular bedding vein impregnating a thin layer of Batatal schist between the Caraça quartzite and the Itabira iron formation. In the old workings near Ouro Preto quartz veins are found cutting the Caraça quartzite and the Itabira iron formation. Near Cattas Altas old workings are found in the basement complex and here also quartz veins occur. Many other examples might be given of quartz veins found in the various formations.

The mineral veins of the district for the most part come under two general heads: (1) ordinary quartz veins, and (2) veins of magmatic origin. To the first class belong certain quartz-hematite veins occurring in the iron formation, and the numerous barren

quartz veins found throughout the district, while to the second class belong the quartz-feldspar pegmatites and the gold-bearing quartz, or quartz-sulphide veins.

The quartz-hematite veins in the iron formation consist of quartz in which flakes of specular hematite occur. The hematite may be found in large, irregular, curved flakes along cracks, or it may be imbedded within the solid quartz. These and various barren quartz veins are of little or no importance as gold-bearing veins. They are probably ordinary circulating water depositions, as the quartz does not contain inclusions of monazite, zircon, rutile, garnet, or xenotime, some of which are nearly always present in the quartz in pegmatite veins and in igneous rocks.[1] It is possible, however, that this evidence may not be conclusive and that by more detailed study some of these veins may be found to be hot-water depositions.

The quartz-feldspar pegmatite veins have already been mentioned in the description of the rocks of the basement complex. They consist of quartz, feldspar, and muscovite with which other minerals, such as beryl, columbite, tourmaline, etc., are sometimes associated in minor quantities. They occur only in rocks of the basement complex and are not gold-bearing.

The gold-bearing quartz and sulphide veins are the important sources of gold in the district. They occur in various formations and differ somewhat in mineral composition in different localities. In some places free gold occurs in quartz, sulphides being inconspicuous, while elsewhere the gold occurs in arsenopyrite, pyrrhotite, or pyrite associated with a variable amount of quartz. Other minerals occurring in places in or near these veins are calcite, cyanite, biotite, garnet, oligoclase, tourmaline, albite, siderite, muscovite, and others.

At the Passagem mine[2] the principal sulphide minerals are arsenopyrite, pyrrhotite, and pyrite, occurring in a gangue of quartz, or quartz and decomposed oligoclase, strongly impregnated with tourmaline. The oligoclase is altered to calcite and white mica. With these occur in varying abundance the other minerals

[1] Specimens examined by Dr. Derby.

[2] A. J. Bensusan, "The Passagem Mine and Works," *Inst. Min. and Met.*, Twentieth Session, October 19, 1910.

mentioned above. Dr. Derby[1] has made a study of the genesis of the mineral and ore deposition at this mine and has established three successive stages of mineralization. He believes that the original quartz-oligoclase deposition was of a pegmatitic nature, the material being derived from some intrusive igneous mass which, however, has not yet been discovered. For although nowhere in the district has any evidence been found of igneous intrusions into the sedimentary series, the nature of the mineralization is such as to leave no doubt as to its magmatic origin. During the deposition of the quartz and oligoclase it appears that garnet (andradite), biotite, cyanite and staurolite were developed locally along the contact, these minerals occurring in the country rock along the border of the vein. Crystals of apatite are associated sparingly with the garnet. Graphite also is associated with the contact minerals, being found along shearing planes.

The second stage of mineralization, according to Dr. Derby, consisted in the introduction of tourmaline along cracks and as impregnations into the earlier minerals.[2] This stage was very pronounced, tourmaline being one of the principal gangue minerals. Its introduction was closely followed by the third stage, namely the sulphide mineralization, by which were introduced arsenopyrite, pyrrhotite, and pyrite. During the introduction of both the tourmaline and sulphides, the alteration of the oligoclase to calcite and white mica took place. Both tourmaline and sulphides occur with the white mica and calcite in the decomposed oligoclase masses, their abundance depending on the degree of alteration. There is evidence that the tourmaline and sulphide solutions attacked the contact minerals as well as the vein material. The biotite is decomposed in places, and tourmaline and sulphides are commonly associated with the contact minerals.

In the vicinity of the gold-bearing lode, but not directly connected with it, are certain geodes with the following mineral association: calcite, siderite, albite, quartz and muscovite. The

[1] O. A. Derby, "On the Mineralization of the Gold Bearing Lode of Passagem, Minas Geraes, Brazil," *Am. Jour. Sci.*, XXXII (1911), 185–90.

[2] *Op. cit.*, p. 190.

origin of these and their relation to the minerals of the gold-bearing lode has not yet been determined.

Summarizing the foregoing statements we may separate the various minerals into the following groups:

Original Vein Minerals	Contact Minerals	Later Vein Minerals		Geode Minerals
Quartz	Garnet	Tourmaline	Arsenopyrite	Calcite
Oligoclase	Biotite		Pyrrhotite	(surface etched)
(largely altered	(locally		Pyrite	Siderite
to white mica	altered)			(surface dissolved)
and calcite)	Staurolite	Secondary Minerals		Albite
	Cyanite	Sericite		Quartz
	Apatite	Calcite		Muscovite
	Graphite			(with sericite)
	(associated			
	with contact			
	minerals)			

The famous Morro Velho mine at Villa Nova de Lima, which has been the greatest gold-producer in Minas Geraes, is operating on a vein in the Piracicaba schist. This vein is parallel to the bedding which here dips toward the southeast at an angle of about 45°. It has an average width of 5 to 10 meters, and this notable width continues with some variation to the greatest depth yet reached, which in 1912 was more than 1,600 meters below the surface. The lode reaches a length of more than 150 meters along the strike.

While the country rock is the Piracicaba schist, the gangue material is a fine-textured mixture of carbonates (siderite, dolomite, and calcite) with quartz, and, according to Dr. Derby, a small amount of albite.[1] The ore consists principally of pyrrhotite with smaller amounts of arsenopyrite, pyrite, and chalcopyrite. A notable feature of the vein is its constancy in width, mineralization, and values through the numerous levels down to the lowest one yet opened.[2]

[1] O. A. Derby, "Notes on Brazilian Gold Ores," *Trans. Am. Inst. Mining Eng.*, XXXIII (1902), p. 282–87.

[2] For a brief history and general statistics of these mines see: H. K. Scott, "The Gold-field of the State of Minas Geraes, Brazil," *Trans. Am. Inst. Min. Eng.*, XXXIII (1902), 406–44.

The gold disseminated through the iron formation and resulting canga formed the basis for most of the gold-mining operations in the early years. A considerable proportion of the iron formation being soft, it was washed almost as easily as stream gravels and, therefore, was attractive to the early miners.

Gold occurs as the native element disseminated in the Itabira iron formation, as well as in the iron-formation lenses of the Piracicaba schist. It may be found in hard, or soft ore, or in itabirite, but generally it occurs in the soft schistose phase of the iron formation known as "jacutinga" (see p. 389).[1] Most of the old workings are in soft itabirite, or in soft, powdery ore, doubtless because of the facility of operating in these formations. The gold is very unequally disseminated within the iron-formation belts, there being large areas of iron formation in which little or no gold occurs, and other places where it is found gathered in rich pockets. There is no regularity in the distribution of the gold-bearing portions of the iron formation and no apparent physical condition, unless it be difference in porosity, which would cause the localization in the places where it occurs. Because of the irregularity of distribution and the disseminated nature of the gold, this type of deposits has never formed the basis of large individual operations. The deposits were worked during the time of slavery because of the cheapness of labor, but when slavery was abolished the workings were abandoned.

The gold occurring in the iron formation is of varying coarseness, from fine dust, hardly visible to the naked eye, to fragments several centimeters in length. Even large masses are reported to have been found. The larger pieces have a porous, spongy texture and contain intermixed particles of iron oxide and quartz. The particles of gold are generally elongated or platy in form, owing to deposition along the lamination planes of the iron formation.

Most of the gold occurring in the iron formation is very pure and contains but very little alloyed silver or other metals. Locally, however, there are occurrences of what is known as *ouro branco* (white gold), which consists of an alloy of gold and palladium.

[1] E. Hussak, "O Palladio e a Platina no Brazil," *Annaes da Escola de Minas de Ouro Preto*, N. 8 (1906), p. 96, tr. by Miguel A. R. Lisboa, and Manoel A. R. Lisboa.

This alloy has been encountered in a number of places, but is nowhere very abundant.

Quartz in veins or lenses is very common in the iron formation, and in places contains gold. It is very often found where disseminated gold occurs, and it is possible that some genetic relation exists between the gold in the quartz and that disseminated in the iron formation. As the masses of vein quartz in this association are generally small, irregular, and discontinuous, it seems possible that, because of the porous nature of the formation, the gold-bearing waters did not follow regular channels but impregnated the iron formation for considerable distances on both sides of the main lines of flow and thus deposited their gold in a more or less disseminated condition. At any rate, from its form and occurrence it is certain that the gold is a secondary concentration within the iron formation.

Gold occurs in the gravels of most streams rising in, or flowing through, the gold-bearing district (Fig. 19). These occurrences are simply placer deposits whose gold has been derived in part from gold-bearing quartz veins, but mainly from the great masses of iron formation, the disseminated gold of which, because of its widespread occurrence, offers abundant opportunity for concentration along streams.

At present small gold-washing operations are conducted by natives along many of the streams at the end of each rainy season, since the flood waters caused by the heavy tropical rains bring new material down the valleys each year.

DIAMONDS

The principal diamond fields of Brazil are those of central Minas Geraes near Diamantina, of Western Minas Geraes near Bagagem, and of Bahia near the towns of Sincora, Lencoēs, and Jacobina. Of these the Diamantina field is the oldest and best known and has yielded most of the best Brazilian diamonds. The Bagagem field has yielded most of the large diamonds found in Brazil, while the Bahia field supplies nearly all the world's demand for carbonados.

The Diamantina field is located in the Serra do Espinhaço quartzite belt, the principal workings lying north, east, and south

of the town of Diamantina. They occur on portions of the main tableland which forms the watershed between the Rio Jequitinhonha and the Rio São Francisco, as well as on highlands between the upper tributaries of the Rio Jequitinhonha.

The Serro do Espinhaço in this portion of its extent is a belt of uneven, rolling tableland from which rise numerous small, irregular quartzite knobs and ridges. The plateau is deeply cut into by the

FIG. 19.—The Rio Gualaxo at Antonio Pereira. The scene of much placer mining in the past and where small gold-washing operations are still conducted by natives at the end of each rainy season. The prominent peak is that of Frazão, a lens of hard quartzite in the Piracicaba schist.

headwaters of the Jequitinhonha which form great, rocky gorges. The gorges are, however, comparatively recent, and by far the larger portion of the region still remains as broad stretches of uplands known in Brazil as *chapadas*. Above the general level of the uplands in the eastern part of the district rises an old monadnock, the isolated peak of Itambé, which is one of the highest points in Minas Geraes. The uplands have a general elevation varying from

1,200 meters to 1,300 meters above sea-level. The Jequitinhonha where it leaves the district has an elevation of about 900 meters, while Itambé rises above the uplands to an elevation of 2,000 meters above the sea; thus the greatest relief in the district is about 1,100 meters.

The tablelands or *chapadas* are of great interest, both geologically and commercially, for it was on them that the diamonds were

FIG. 20.—The diamond-washing of Serrinho near Curralinho. In the distance is the monadnock of Itambé.

first concentrated and it is on them that the principal deposits are now found. The *chapadas* are, for the most part, underlain by Caraça quartzite, dipping uniformly but generally at low angles to the east. On many of them, especially those near the actual watershed of the Serra do Espinhaço, there are remnants of what were once more extensive deposits of gravel, sand, and clay. These have been called the Diamantina conglomerate. The remnants lie on the eroded surface of the Caraça quartzite, and when examined carefully it is found that most of them occur in irregular trough-like depressions, the bottom and walls of which consist

of quartzite (Fig. 21). Many of these troughs reach a depth of 30 meters or more, though some of the deeper ones scarcely exceed 100 meters in width (Fig. 22). No determination has been made of their maximum width, while as to their original longitudinal extent nothing is known. From their irregular occurrence and their relation to the underlying quartzite it is presumed that these deposits are for the most part the result of stream deposition.

FIG. 21.—The pit of Serrinho. A shallow depression in the Caraça quartzite which was partially filled with diamond-bearing conglomerate. The conglomerate and beautifully white kaolin are seen in the bottom of the excavation. The final washing operations to recover the diamonds are conducted in the *batêa*, or jig, under the thatched roof.

This view is further strengthened by the nature of the materials composing them and the relation of the different kinds of material to each other. By detailed mapping it might be possible to connect all these remnants so as to reproduce the old drainage network as it was when the conglomerate was deposited.

The gravel, sand, and clay occur associated in the deposits as beds, lenses, or masses, without any regularity. Locally beds of pure

clay, or clay slightly intermixed with sand, occur, while elsewhere abundant pebbles are scattered through clay or sand matrices. The nature of the pebbles in different parts of the district varies, but the characteristic and dominant pebbles consist of quartzite, undoubtedly derived from the Caraça formation. Associated with these are pebbles of iron formation, quartz, schist, diorite, amphibolite, and other rocks, the association varying in different

FIG. 22.—The trench at Cadette's mine, northwest of São João da Chapada

localities. A characteristic but less abundant pebble in these deposits is the diamond.

Where exposed at the surface the conglomerate is very hard and indurated, but beneath the surface both matrix and pebbles, especially those of quartzite, are in many places soft and friable. This softening of the quartzite pebbles bears evidence of considerable age of the conglomerate deposits.

In many respects these deposits are strikingly similar to certain deposits of gravel, sand, and clay which occur along the Appalachian region in the United States. In Georgia these contain bauxite;

farther north in scattered localities they contain concentration deposits of brown iron ore and manganese ore. In the United States these deposits have been referred to the Tertiary or late Cretaceous.

From the great predominance of quartzite pebbles it must be supposed that the materials composing these soft conglomerate deposits came largely from the Caraça quartzite. The immediate source of the igneous pebbles present in some localities, however, is less certain. These may once have been laid down in conglomerate beds in the Caraça quartzite and have been worked over again later, or they may have been derived from the rocks of the igneous area adjacent to the Espinhaço sedimentary belt. The Caraça formation in this district is a well-consolidated quartzite of medium-grained uniform texture. Conglomerate beds are rare, and though scattered pebbles are frequent, these are almost invariably of vein quartz. There is a possibility, however, that pebbles of igneous rock may be present locally.

Accepting the supposition that the material composing the conglomerate was derived largely from the Caraça quartzite, one would naturally look for the origin of the diamonds in the same formation. In the Diamantina district, diamonds have never to our knowledge been found in the Caraça quartzite. It is, however, possible that they may occur so widely scattered that it is only by repeated concentration such as has occurred in the conglomerate deposits that they become noticeable. In the Grão Mogul district of northern Minas Geraes about 200 kilometers to the northeast of Diamantina, diamonds are reported to be found in hard quartzite[1] from which they are freed by blasting. However, it is not known whether this is the Caraça quartzite or whether it belongs to some other formation.

In conformity with the generally accepted theory of their origin, the diamonds may have been derived originally from intrusions of igneous rock at a distance from the present areas of concentration. Disintegration and decomposition of the inclosing rocks would have resulted in the freeing of the diamonds, leaving them

[1] O. A. Derby, "Modes of Occurrence of the Diamond in Brazil," *Am. Jour. Sci.*, 3d Ser., XXIV, 39.

in residual accumulations subject to removal by streams and other subaerial agencies. If in part transported to the sea during the deposition of the Caraça formation, they should be found probably in association with quartz sand and such pebbles of vein quartz as resisted the abrasian of stream action. Being the hardest of all minerals and the most resistant to decomposition, it is not surprising that the diamonds still retain their crystalline faces when most other minerals originally present, except quartz, have disappeared.

When in pre-Devonian times the sedimentary series was elevated, consolidated, and metamorphosed, and the processes of erosion had commenced, the diamonds began gradually one by one to be freed from the quartzite and collected in the stream deposits along with other minerals from the same source. This continued for long ages, and while other minerals were disintegrating and decomposing, the diamonds remained intact and became more and more concentrated. Even after a general peneplain level was reached, this process probably continued and was only interrupted by renewed elevation resulting in the present gorges.

The foregoing discussion is based on the hypothesis that the diamonds at one stage in their history were incorporated in the Caraça quartzite. The alternative hypothesis is that the majority of them never have been deposited in the quartzite, but that they were brought in by streams directly from igneous areas some distance from the present diamond fields. An objection to this hypothesis is that so few igneous pebbles occur in the conglomerate, and that over large areas pebbles of igneous rocks are entirely absent. However, it is possible that, on account of their predisposition to rapid decay, they might have disappeared in the slow, shifting process on the peneplain surface. But one wonders why, if such diamond-bearing igneous-rock areas exist, not concealed by the sedimentary rocks, they have not been discovered up to the present time. If the diamonds were brought in from a long distance the *chapada* diamond-bearing gravels should necessarily occur along a few main drainage courses and not be irregularly scattered over a wide area, while if they came from igneous areas close by, the diamonds should have been discovered in streams flowing from these areas. But in the present state of knowledge it

seems best to leave the question whether the diamonds in these *chapada* conglomerate deposits have, for the most part, been derived directly from igneous rocks of the Archean complex, or whether, coming in any case from that source originally, there occurred an intermediate stage of incorporation in the Caraça quartzite—an open one.

Perhaps the most noted of the diamond mines of this district is that of São João da Chapada, which has been rather fully described by Dr. Derby.[1] It is located some 30 kilometers northwest of Diamantina on the tableland which represents the old base-leveled surface. Sunk in the solid quartzite of the tableland is a steep-walled trench, perhaps 500 meters long and now about 30 meters in depth (Fig. 23). This trench has the appearance of being the channel of an old stream which became completely filled with residual clay, sand, and gravel when the surrounding region was close to the base level. This alluvial-filled trench has been cleaned out again as the deposits of clay, sand, and gravel have been washed away in the process of diamond mining.

Dr. Derby in several of his papers has been inclined to the hypothesis that the diamonds of São João da Chapada are vein minerals,[2] derived from a pegmatite vein of which there seems to be some evidence in the cut. Certain masses of pure-white kaolin containing nests of large and beautiful crystals of quartz, which have never been exposed to the wear of running water, look much like the decay products of a pegmatite vein. Unfortunately at the time of our visit in 1912 the mine had not been worked for twenty years, so that definite and reliable information was not easily obtained, but from the accounts given by some of the people at São João da Chapada and the published papers which touch upon this point, it would seem that the diamonds were far more abundant and characteristic of the waterworn conglomerates, especially where pebbles of higher specific gravity (iron ore, itabirite, etc.) were

[1] O. A. Derby, "Brazilian Evidence on the Genesis of the Diamond," *Jour. Geol.* VI (1898), 121–46; "On the Association of Argillaceous Rocks with Quartz Veins in the Region of Diamantina, Brazil," *Am. Jour. Sci.*, 4th Ser., VII 343–56.

[2] O. A. Derby, "Modes of Occurrence of the Diamond in Brazil," *Am. Jour. Sci.* 3d Ser., XXIV, 34–42; "The Genesis of the Diamond," *Science*, IX (1887), 57–58.

common, than of those particular clayey portions which suggest a pegmatitic source. Because of this the São João da Chapada occurrence does not seem essentially different from the other diamond-bearing conglomerate deposits. A pegmatitic dike, being less resistant than the surrounding quartzite, may have determined the location of the drainage channel in the first place, and because of its ready yielding gave rise to the steep-sided trench which would

FIG. 23.—The famous diamond mine of São João da Chapada. The conglomerate-filled trench in the quartzite has now been largely re-excavated, the undisturbed quartzite appearing at various points in both walls and in the bottom of the excavation. This mine has produced some of the finest of the Brazilian diamonds.

otherwise seem peculiar in a peneplained surface. But we believe that if such pegmatitic material was present it was only a contributing condition, and that the diamonds came with the material that was washed into the trench filling it up.

Since their deposition with the residual base-level gravels which now constitute the *chapada* or upland deposits, several reconcentrations of the diamonds have occurred, as the region has suffered various changes of attitude. Upon the rejuvenation of the drainage

the streams began dissecting the old nearly base-leveled plain, first cutting through the surface decomposition products and associated fluviatile deposits, and then attacking the underlying quartzite. As the *chapada* deposits were being removed in this erosive process, the softer and lighter materials were being worn finer and washed away, while the heavier diamonds remained behind and became further concentrated in the stream gravels of the immediate

FIG. 24.—Diamond mining at Cadette's mine. The clay and materials of lesser specific gravity are washed away by running water while the diamonds and other heavy pebbles remain behind and become concentrated.

vicinity. At several stages during this erosion there were periods when the downward cutting of the streams was checked and they began to widen their valley bottoms and deposit material over them. Then active cutting began again, leaving gravel-covered terraces on the valley slopes. Much material from these several gravel terraces as well as from the diamondiferous conglomerate of high-level *chapadas* has since been washed down the present steep

valley slope into the Jequitinhonha River. This has resulted in the following distribution of the diamond-bearing deposits. Numerous remnants of *chapada* deposits still occur upon the plateau in areas as yet undissected by the rejuvenated streams. Gravel deposits of later age occur lower down on valley terraces at various elevations above the present stream beds. Gravel and sand deposits of still more recent age occur along the present stream

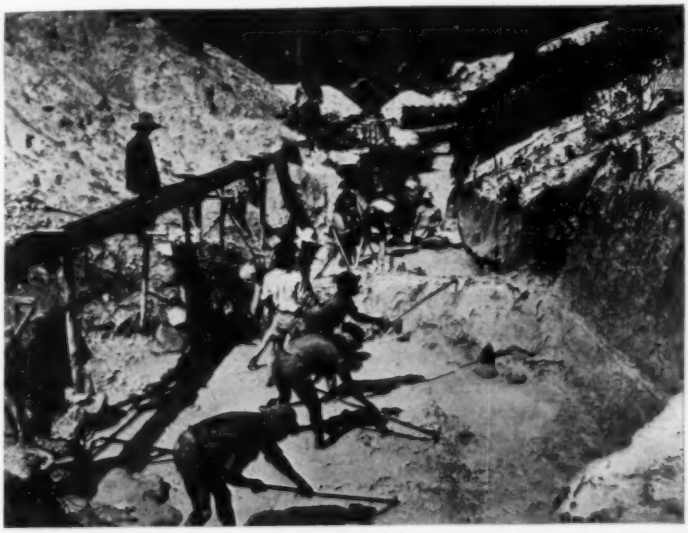

FIG. 25.—Stirring up the mortar-like mass of diamond-bearing clay and gravel

bottoms. All these deposits contain diamonds in greater or less abundance.

The present-day practice for the recovery of the diamonds is simply a furtherance of the process of concentration which has been going on in nature. When a considerable quantity of soft kaolin and loose gravel from the diamondiferous conglomeratic deposits has been accumulated in the bottom of the trench, or mine, a stream of water is conducted through the trench (Fig. 24). As the running water passes over this loose material, natives armed with a sort of

hoe keep the mass constantly stirred up, so that it has a general consistency and appearance not unlike thin mortar (Fig. 25). The running water carries away the clay and materials of lesser specific gravity, while the coarse portions of the gravel, and the minerals of higher specific gravity such as the diamond, fragments of hematite, etc., remain behind. Later, after screening out the coarser pebbles, the diamonds are picked out of the final concentrate on the *batêa*, or washing sieve, by hand.

The diamonds of the Diamantina district generally show dodecahedral crystallization with rounded faces. Many of those in the *chapada* deposits have rough faces and a dark grayish-green coating, but the majority have smooth or striated faces. Some of the stones in the present river gravels show surface corrasion.

The majority of the diamonds mined in the district vary in size from $\frac{1}{4}$ carat to 4 or 5 carats, though occasionally stones are found weighing up to 10 carats. The largest diamonds ever found in the Diamantina district, according to Dr. Derby,[1] weighed less than 100 carats, and very few have been found weighing over 50 carats. The largest of the Brazilian diamonds—the "Star of the South," the "Star of Minas," and the "Dresden" diamonds—have all come from the Bagagem district in southwestern Minas Geraes.

In brilliancy the Diamantina diamonds exceed the South African diamonds, and there is a smaller percentage of "off-colored" stones found. Beautiful bluish-white stones are abundant, and stones of other colors such as lemon-yellow, cognac-brown, rose, green, and blue are occasionally found.

We are indebted to Professor T. C. Chamberlin, Professor C. K. Leith, and Professor Eliot Blackwelder for criticisms and suggestions, and to Dr. O. A. Derby for much information.

[1] O. A. Derby, "A Notable Brazilian Diamond," *Am. Jour. Sci.*, XXXII (1911), 191.

THE STRENGTH OF THE EARTH'S CRUST

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PART VIII. PHYSICAL CONDITIONS CONTROLLING THE NATURE OF LITHOSPHERE AND ASTHENOSPHERE

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RELATIONS BETWEEN RIGIDITY, STRENGTH, AND IGNEOUS ACTIVITY

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INTRODUCTION AND SUMMARY

The experiments of F. D. Adams have demonstrated that under combined pressures and temperatures equal to those existing at a depth of eleven miles granite is about seven times stronger than at the surface, showing that, up to at least certain limits, the strength of the crust increases downward. The measurements of tidal deformation of the earth and of the variations of latitude concur furthermore, in proving that the rigidity of the earth as a whole is greater than that of steel. The transmission of earthquake vibrations of a transverse nature through the earth shows, not only that the earth is solid and rigid throughout, but that, on the whole, rigidity increases with depth. In none of these lines of investigation is there any clear suggestion of the existence of a thick shell of weakness—an asthenosphere.

On the other hand, the conclusion that broad areas of the crust rest in approximate isostatic equilibrium seems to imply the

[1] Section B of Part VIII, on "Relations with Other Fields of Geophysics," will be published in the succeeding number of this *Journal*.

existence of a subcrustal zone with but little strength, readily yielding under vertical loads when these are of such breadth that the strains resulting from them are not confined and absorbed within the strong outer crust.

There has thus been developed a paradox, an apparent conflict of evidence which becomes more insistent of explanation with continued accumulation of proofs of high rigidity from the domain of geophysics and of proofs of regional isostasy from the equally precise field of geodesy.

In the consideration of such broad problems, the hope of an ultimate definitive solution rests upon the use of the method of multiple working hypotheses. The surety, significance, and breadth of application of the facts must be established. By these the various hypotheses must be tested and molded. All hypotheses must be kept for further consideration provided they are not positively excluded. In the complexity of relationships there is commonly a complexity of cause, and hypotheses which seem at first to be mutually exclusive may be found to co-operate in giving a completer explanation. Those which originally appeared antagonistic may thus come to be seen as participating and dividing the field of cause between them. A paradox often points to this kind of a conclusion.

To pass to the particular problem of the relations of isostasy to the physical conditions of the earth's interior; the hypothesis developed in this study—of the existence of a zone of weakness underlying a zone of strength—must not be regarded at present as the only available hypothesis. Searching investigation must be carried forward to see if other and possibly antagonistic hypotheses cannot be developed which will equally well co-ordinate and explain the facts. Even if true in the main, it is likely, as has been the case with other hypotheses, that time will show that in certain directions it has been carried too far. Such testing, however, can best be done by others, and after the implications of this hypothesis are seen. In this part, the concluding article of this series, a discussion had best be given of the lines of adjustment by which the hypothesis here favored may be brought into harmony with other fields of geophysical evidence. With this understanding of the relation of

the present investigation to the method of multiple working hypotheses, examination will be made of the paradox which has been drawn between certain conceptions from other lines of investigation and those drawn from this study of crustal strength.

Having given this introductory presentation on what is conceived to be a judicial point of view, we may turn to a review of the conclusions reached in this article. It is pointed out that rigidity is strictly a measure of stiffness; whereas a very different quality, the limit of elastic yielding, or the beginning of flow, is the measure of strength. But mass flowage may take place in a number of quite different ways, according to the nature of the solid and the environing physical and chemical conditions. The elastic limit and hence the strength will differ in the same solid according to the mode of yielding. Four modes may be here enumerated in what is thought to be their order of increasing importance, the fourth mode being that which is conceived as operative especially in the asthenosphere, and serving to maintain the condition of approximate regional isostasy.

First, flowage may take place rapidly by true plastic or molecular flow, as with lead or white-hot iron, the solid, when stressed well beyond the elastic limit, behaving like a viscous fluid. It is not thought that the terrestrial deformations are often carried on with a rapidity which requires true plastic yielding. In fact, under such rapid stresses as those produced by earthquakes and tides it is not improbable that the strength of the earth may progressively increase with depth.

As a second and quite different mode, deformation may take place by molar as distinct from molecular shear. In the zone of fracture this is manifested in jointing and faulting and is emphasized as distinct from rock flowage, but where the fracturing becomes so closely spaced as to result in slicing of individual minerals it passes under the category of granulation. Where carried on at depth there is always some degree of cementation by recrystallization. Deformation by such close-grained fracture without complete loss of cohesion is classed as rock flowage. It is thought to be developed to some degree within the lithosphere, especially by great horizontally compressive forces, but is not

regarded here as the mode of deep rock flowage involved in the isostatic readjustment of unfolded tracts.

Thirdly, flowage may take place in some minerals, as calcite and ice, by gliding upon the cleavage planes. But such gliding is not regarded as the mode by which the foliated rocks are developed. It requires furthermore far greater force than that which is given by the departures from isostatic equilibrium.

A fourth mode of rock flowage is by recrystallization. It is the chief factor, as Van Hise has shown, in the deformation of the crystalline foliates. It is thought that this is also the method by which the asthenosphere yields and that a readiness of recrystallization under unbalanced stresses of a permanent nature is the cause of the weakness of the asthenosphere.

The vibratory forces transmitted as earthquake shocks and those due to tidal strain, from this standpoint, are both rapid. Under such conditions the asthenosphere could show high order of strength. It is argued that the ease of recrystallization under constant strain becomes more marked the nearer the temperature approaches to that of fusion, or to express it better from the physico-chemical standpoint, the nearer the temperature approaches to the mutual solution point of the constitutents involved. The result is that at such temperatures the rigidity may be high and not greatly different from that at low temperatures, but for permanent stresses the elastic limit becomes low. The movement of continental glaciers with a low surface gradient, accomplished by recrystallization, illustrates the condition which it would appear exists to even a higher degree within the asthenosphere.

This conclusion carries with it the idea that within the lithosphere the temperature is in general considerably below that of fusion; whereas below, in the thick zone of weakness, the temperature must lie close to, or at, that of molten rocks. For fusion there is needed, however, the energy necessary to supply the latent heat and volume expansion. Unless this energy is supplied, the asthenosphere remains solid rock, but the least accession of internal heat, or relief from external pressure, will generate a proportionate amount of magma, diffused as liquid throughout the solid. To gather into reservoirs temporarily molten, the magma must con-

verge by rising, analogous to the draining of melted water from glaciers; uniting, as rivulets unite into rivers, and rivers discharge into lakes. No continuous lava stratum or large reservoirs of lava could, under the terms of this hypothesis, be expected to exist within the asthenosphere. Its very weakness would prevent it from acting as a containing vessel for holding large volumes of any fluid which, for any cause such as a lower specific gravity of the fluid phase, would tend to rise. The evidence of earthquake vibrations and of resistance to tidal deformation further supports the view that the asthenosphere is not a liquid or even a truly viscous zone. On the other hand, only in the lithosphere would be found the strength needed for the storage of magma in volumes until the limit of its strength as a containing vessel was reached.

Partly guided by observation upon the metamorphic rocks, partly by theories of the nature of deformation at great depths, the argument leads to conclusions on the mode of yielding within the different levels of the crust. First, the outermost zone is observed to be a zone of fracture, weak in comparison with the thick zone below. This, the second zone, is the zone of strength and yields by flowage, but flowage which is characterized by granulation as the dominant, by recrystallization as the subordinate, mode. The expenditure of energy for a given deformation is here a maximum. In the third zone, the asthenosphere, on the contrary, flowage is conceived as taking place with but little expenditure of energy, by a ready recrystallization at the temperature of primary crystallization of magmas. Those contorted granite-gneisses seen especially in the Archean rocks, which are regarded as deformed during the final stages of crystallization, exhibit locally in the outer crust the conditions which it appears may permanently prevail within the asthenosphere.

SECTION A

RELATIONS BETWEEN RIGIDITY, STRENGTH AND IGNEOUS ACTIVITY

DISTINCTIONS IN PHYSICAL PROPERTIES RELATED TO STRENGTH

Elasticity is of two natures: that of volume and that of form. The first is possessed by matter in either the gaseous, liquid, or solid state; the second is possessed by solids only and is associated

with rigidity and strength. Up to a certain degree of strain known as the elastic limit, elasticity of form in the ideal solid is perfect and is expressed by the law that the change of form, or strain, is directly proportional to the load applied, or stress. This load may be maintained indefinitely and, except for a slight relaxation, the solid shows no further yielding. Upon the removal of the load there is an elastic return to the original form, but the very last stages of the recovery are slow. The elastic nature of the whole earth, in regard to both volume and form, is shown by its capacity to transmit the several kinds of earthquake vibrations. The permanence and perfection of the elasticity of form is also implied in the power of the crust to carry loads up to certain limits for times reaching into geologic periods without exhibiting progressive viscous yielding.

Beyond the elastic limit, elasticity ceases to be perfect, and a permanent change of form occurs. This relieves part of the stress and reduces the strain to within the elastic limit. The change of form may be by rupture, in which event the strength of the body is destroyed. It may be by plastic flow, in which case the strength may be increased or decreased. Wrought iron, for example, becomes somewhat stronger as a result of forging. Granite on being mashed into gneiss becomes somewhat weaker because of the development of weaker minerals, especially the micas. In the crust of the earth, except for the outer few miles, flowage takes place without probably much change in the mineral composition and consequently in the strength of the rock. Deformation will continue as long as the stress is maintained well above the elastic limit, but upon the cessation of movement there may still remain residual stresses up to the elastic limit. If the residual stresses over broad areas are small, it may be because the development of weaker structures, such as folds or zones of igneous injection, has eased the strains.

Failure by flow brings in the distinction between viscosity and plasticity. These are often used, even by physical geologists, as merely synonymous terms, but there is a real distinction which should be noted. Fluids are viscous to a small or large degree and can have no elasticity of form. Viscous flow must, however, overcome internal friction and requires time for its accomplishment.

With long time even a minute force will cause even a very viscous fluid to flow. Solids, on the other hand, possess elasticity of form, and below the elastic limit can hold shearing stresses indefinitely. Above it they may flow and in so doing exhibit plasticity. The phenomenon differs from viscosity in that the force must rise to a certain magnitude before any gliding between molecules begins. The crust, then, is plastic but not viscous.

Although the theoretical distinction between plasticity and viscosity is clear, recognition must be given to conditions where the two states merge. This is especially true for undercooled glasses. A glass in its molecular organization is a liquid and yet it possesses definite elastic moduli and elastic limit. From this standpoint of elasticity the glass, therefore, is a solid. Upon rise of temperature there is, however, no absorption of latent heat to mark a change of state, the elastic limit gradually lowers, disappears for prolonged stresses, and elastically, the substance passes by gradation from a solid to a liquid. The existence of these transition cases should not, however, be permitted to obscure the real distinctions between solid and liquid.

The crust yields as a plastic solid to forces which strain it beyond its elastic limit. But the solid flowage which this implies may be either by distortion of crystals or by recrystallization. The first is familiar for rapidly applied forces, requires comparatively great stress, and corresponds to the usual conception of plasticity. The crystalline rocks make us familiar, however, with the idea of mass plasticity by recrystallization. This is plasticity, but in a somewhat different sense from that which is usually conveyed by the term.

The degree of elasticity which a substance may exhibit is a different property from the elastic limit. A bar of wrought iron one square inch in cross-section will be elongated one part in 28,000,000 by a tensile stress of one pound. A similar bar of glass would be elongated one part in 10,500,000, more than twice as much. These ratios measure the degree of elasticity under tensile or compressive stresses and differ for each substance. The figure is known as Young's modulus of elasticity. A substance may be highly elastic, that is, have a high modulus of elasticity, as cast iron, or glass,

and yet be brittle because of a low elastic limit under rapid tensile stress, combined with lack of plasticity at ordinary temperatures. At temperatures sufficiently high, the modulus is not greatly different, but the elastic limit is still lower. The substance is now, however, plastic, rather than brittle, since plasticity is greatly increased; but a rapid strain, exceeding the rapidity with which plastic deformation can take place, may still produce fracture. Another substance, such as rubber, may have a low modulus of elasticity and yet a relatively high elastic limit.

Among similar substances under similar physical conditions there is, however, a definite association of these properties which for the metals is brought out well in a tabulation by Johnston and L. H. Adams.[1] It is shown for a class of substances, such as the metals, that the modulus of elasticity, the hardness, the tensile strength, and the elastic limit all, so far as the data are given, occur in the same order; so that of two metals that which has the higher elastic limit is the higher also in the other qualities. From this association there results a ready mental confusion between rigidity and strength. The one, however, denotes the degree of resistance to distortion from a unit-shearing stress and gives the modulus of ridigity. The other is measured by the elastic limit. As an example of the confusion between these two different properties, it is known that the earth as a whole is more rigid than steel. This to many would appear to mean that it was stronger than steel. Earthquake waves show that the earth becomes progressively more incompressible and more rigid with depth. This might be held as evidence against the existence of a thick sphere of weakness, the asthenosphere. High incompressibility and high rigidity are not, however, direct testimony of strength, and it is the purpose of the next topic to show under what conditions a solid may be very rigid and yet very weak.

CONDITIONS FAVORING ASSOCIATION OF HIGH RIGIDITY WITH LOW ELASTIC LIMIT

Alpine glaciers as well as the Alpine-like margins of the Greenland ice sheet move much more rapidly in the summer than in the

[1] "On the Effect of High Pressures on the Physical and Chemical Behavior of Solids," *Amer. Jour. Sci.*, XXXV (1913), 220.

winter, a phenomenon to be accounted for by the rate of recrystallization. The parts of an ice crystal which are subjected to shear and compression have the melting-point lowered. They melt, discharge the strain, and refreeze. In the winter the general temperature is reduced, and a greater strain is necessary to bring the melting-point down to the lower temperature. Until local melting is produced the ice behaves like any other crystalline solid, as a substance possessing elasticity of form. Beyond that point it exhibits plasticity and behaves in some respects like a very viscous fluid. In other respects, however, it exhibits properties quite distinct from that of the usual conception of mere plastic flow, since in the testing machine, or on the walls of a crevasse, ice will resist strong shearing strains, and yet the glacier as a whole yields and flows slowly under a moderate pressure-difference as shown by the low gradient of its upper surface. Glacial motion appears to take place, therefore, by the solution and growth of crystals, not by a true viscous flow. The solid and crystalline nature throughout as opposed to viscous fluidity is furthermore shown, as Chamberlin has noted, in the power of the glacial ice to shove over and abrade its floor and to ride up slopes. Chamberlin adds that a dry glacier is a rigid glacier. A dry glacier is necessarily cold, and a cold glacier is necessarily dry.[1]

With ice subjected to slowly applied forces the elastic limit is consequently dependent upon the point of yielding by recrystallization. We thus see an intimate relationship between temperature and variation in the elastic limit, the elastic limit for ice being greater for low temperatures than for high temperatures. But the modulus of rigidity, on the contrary, measures the elastic change of form for unit-shearing force, change of form not accompanied by crystallization, but marked by a capacity to spring back to the original form upon the removal of the stress.

T. W. Richards, in his studies on the compressibility of solids, notes that they are almost as compressible and voluminous at absolute zero as at ordinary temperatures. Under this conception

[1] "A Contribution to the Theory of Glacial Motion," *Decennial Publications of the University of Chicago*, IX, 203, 204 (1904); Chamberlin and Salisbury, *Geology*, I (1904), 305.

the molecules, taken as equivalent to their spheres of influence, are in actual contact and suffer mutual compression owing to the attraction of cohesion. The influence of heat is relatively unimportant in determining the density of a solid. The atoms are in most cases even more compressed and distorted by the converging force of chemical affinity than are the molecules by cohesion. This corresponds with the fact that substances of small atomic volume are on the whole more incompressible than those of greater atomic volume. For the more incompressible substances also the decrease in compressibility with added load is relatively little, suggesting that they are already greatly compressed by the forces of chemical affinity.[1] The effect of heat serves only to distend slightly the spheres of influence of the atoms, so long as the substance is in the solid state. Rise of temperature to near the melting-point, as long as there is not a softening by the development of incipient liquidity, should, according to these views, change the elastic properties but slightly. The greatly lessened strength of ice near the melting-point, as expressed in the freedom of regelation, is not, following these ideas, connected with the slightly lessened incompressibility and rigidity. For many substances the problem is complicated, however, by changes of molecular state with changes in temperature and pressure. This is especially true of ice when subjected to extreme ranges in temperature and pressure, as has been shown by Bridgeman; for ordinary glacial ice, however, we deal with but a single state.

For ice at a temperature of $-7°.03$ C. the compressibility has been determined for pressures ranging approximately between 100 and 500 atmospheres. It is found to possess, according to Richards and Speyers, about one-fourth of the compressibility of water at neighboring temperatures and about five times the compressibility of glass.[2] But glass possesses a compressibility between that of acidic and basic holocrystalline igneous rocks. Ice may be taken then as about three or four times as compressible as granite.

[1] "The Present Aspect of the Hypothesis of Compressible Atoms," *Am. Chem. Soc. Jour.*, XXXVI (1914), 2417–39.

[2] T. W. Richards and C. L. Speyers, "The Compressibility of Ice," *Am. Chem. Soc. Jour.*, XXXVI (1914), 491–94.

Now the modulus of rigidity is related to the modulus of compressibility by means of a formula involving Poisson's ratio.[1] This ratio varies for each substance, but for rocks, for iron and steel, and probably for ice, it lies between 0.2 and 0.3 in value, so that in general the rigidity of these substances can be judged roughly by their degree of incompressibility. Consequently it is seen that glacier ice at temperatures such as those which prevail in the body of the moving glacier possesses a degree of incompressibility and rigidity which, if these elastic constants were measures of its strength, would make it wholly incapable of motion on such gradients as are observed. This can be made more obvious by some quantitative statements. Granite and similar rocks, for example, can stand permanently in steep cliffs to heights of thousands of feet. They constitute mountain ranges whose height and steepness are limited entirely by the forces of erosion on the one hand and the strength of the asthenosphere on the other. The cliffs could be very much higher and the mountains much more lofty before glacier-like flow at the base of the mountain mass would occur. In fact, with a compressive strength of 25,000 pounds per square inch, a rectangular block of granite could stand as a vertical wall 22,000 feet high, and of indefinite breadth, without yielding of the base. With a sloping face and supported by spurs such as occur in nature, the height of the granite mass could become considerably greater. For parallel mountain ranges of harmonic form and gentle slopes resting upon a foundation whose compressive strength to indefinite depths was 25,000 pounds per square inch, the mountain crests could stand eleven miles above the valley bottoms before the maximum stress-difference would reach this limit. Even then, if the slopes were as low as those of a continental ice sheet, the failure would not take place by flowage of the mountains laterally into the valleys, but by a vertical settling of the mountains and a vertical upwarping of the valleys. The lateral, plastic flow would be at some depth in the earth. If the asthenosphere were indefinitely rigid, granite mountains of sufficiently gentle slope

[1] Let P = Poisson's ratio; C, the modulus of rigidity; D, the modulus of compressibility. Then $C = \frac{3}{2}\,\frac{1-2P}{1+P}\,D$.

could rise to indefinite heights. This is because the depth of maximum stress-difference would lie at about one-sixth of the wave-length below the mean level of the surface. With increasing wave-length the height of the waves could accordingly be greater without increasing the stress-difference at the trough-line of the waves. The gradient would, however, have to become more gentle; in other words, the amplitude would have to increase at a lesser rate than the wave-length. If the strength of ice were measured by its rigidity it could stand permanently in masses one-third or one-fourth as steep and high as these theoretic limits for granite mountains, without failure by plastic flow. Yet, on the contrary, the great ice fields spread out by flowage of their bases, although their surfaces possess very gentle gradients. The distinction between strength and rigidity in the movement of glaciers is thus clear. The strength of glaciers is limited by the amount of the stress-differences needed to produce slow movement by recrystallization.

Johnston and L. H. Adams have applied this theory of yielding, well known as an explanation of glacial motion, to all plastic flow, and argue that even for those substances, such as the metals and rocks, in which cubic compressibility raises the melting-point, shear greatly lowers it for the parts under stress.[1] They argue from a physico-chemical basis that the most plausible explanation for flow in metals is that the shearing strain is great enough on individual points to produce a change of phase of individual molecules from solid to liquid, even at ordinary temperatures.

Apart from theory as to its explanation, the phenomenon of welding of iron shows for high temperatures a low elastic limit and ready passage beyond into plastic flow. For iron and steel, furthermore, the influence of temperature upon the rigidity has been investigated. Pisati gives the following equations in which n is the value of the modulus of rigidity for temperature t.[2] For iron—

$$n_t = 811 \times 10^6 (1 - .000,206t - .000,000,19t^2 + .000,000,001,1t^3),$$

[1] "On the Effect of High Pressures on the Physical and Chemical Behavior of Solids," *Am. Jour. Sci.*, XXXV (1914), 205–53.

[2] *Smithsonian Physical Tables* (1904), p. 76.

for steel—

$$n_t = 829 \times 10^6 (1 - .000{,}187t - .000{,}000{,}59t^2 + .000{,}000{,}000{,}9t^3).$$

This equation for iron gives a minimum modulus at 314° C. equal to 95 per cent of the modulus at 0° C. For steel the minimum value occurs at 342° C. and is 90 per cent of the modulus at zero. At 528 °C. iron has the same modulus as at 0° C. and at 890° C. steel has the same modulus as at zero. Doubtless 1000° C. is above the limits of the data from which these formulas were derived. For this temperature they may consequently give inaccurate results, but it is of interest to note that the curve gives a modulus of rigidity for iron at that temperature 1.7 times that at 0° C. and for steel 1.1 times that at 0° C. The extrapolation prevents attaching quantitative value to these figures, but the qualitative conclusion may be reached that iron and steel at high temperatures do not exhibit less rigidity than they possess at lower temperatures. It is obvious, however, that above a certain temperature the elastic limit becomes very low, as shown by the capacity for forging, and for strains beyond this limit deformation takes place by plastic flow. That it is not merely incipient fusion is suggested by the maintenance of a crystalline condition through the process of deformation. The subject for iron is doubtless complicated by the fact that iron passes through more than one solid molecular state in being heated up to fusion. Presumably then the equation given for the relation of rigidity to temperature can only be a first approximation to the actual changes.

Let the attention be given next to the crystalline rocks which were once deep-seated and, owing to subjacent batholithic invasion, attained their crystallization at exalted temperatures. It is observed that, although the rock masses have been extensively deformed, the individual crystals have regrown during the process so as to possess compact boundaries, and an internal constitution nearly free from strain. The interpretation is that the deformations due to geologic forces were so slow and the rocks were so saturated with crystallizing agents at high temperatures that recrystallization could nearly keep pace with the deformation, even for temperatures below the range of plasticity. As understood by

the students of anamorphism, the process has depended upon a readier solution of the molecules under shearing stress than of those free from such stress—solution carried on by means of the relatively minute proportion of gaseous crystallizers which were present through the rock mass. Such crystallizers doubtless facilitate the process. They form, in fact, solutions with the rock which may be regarded as mixtures with very low fusion points. But, theoretically, as the temperatures approach those of general fusion the need of such crystallizers diminishes. Moderate shearing stresses can thus liquefy the parts upon which they act and a process analogous to glacial motion sets in for solid rock. The strength of highly heated rock appears then dependent upon the amount by which the temperature is below the melting-point. That zone of the earth which is very weak may then be regarded as approximately at the temperature of fusion. To transform the solid into liquid there is needed only the energy required for latent heat and increase of volume. The proportion of liquid which is generated will vary directly with the amount of hydrostatic pressure removed. Magma, consequently, can be generated in this zone more readily than above in the zone of strength; but it will not be in reservoirs; rather will it be in its place of origin disseminated through the rock mass like water standing in a porous sandstone.[1]

ANALOGIES BETWEEN ASTHENOSPHERIC ROCK AND GLACIAL ICE

The theory of the asthenosphere as here presented is seen to have important relations to other branches of geology. The zone of weakness becomes especially the generator of magmas; the

[1] Recently the writer has learned from Mr. Bailey Willis of an unpublished paper which he gave some years ago to the Geological Society of Washington, in which he outlined his views of the nature of crustal thrusts as illustrated by the Appalachians. In that, and more recently, as a result of studies in the Alps and Andes, he has come to entertain the view that the zone of compensation, the lithosphere, shears over the zone below through the agency of molecular or mass fusion. Deep-seated horizontal shear and igneous intrusion he thus holds have important associations with orogenic movements. We have thus arrived independently at somewhat the same view of the nature of the zone of weakness. The part which recrystallization may play in promoting such movement is suggested in his *Research in China*, Vol. II (1907), "Systematic Geology," pp. 130, 131.

pyrosphere has its roots in the asthenosphere. But in the attempt to frame a logical picture of the processes which determine the ascent of magmas, the question arises how diffused liquid matter is drained away, rising and uniting at higher levels into magma reservoirs, temporarily molten, and the direct source of the igneous activity exhibited within the outermost crust. A deductive picture is as follows—one whose truth cannot be tested directly, but only by its general agreement with our understanding.

At the place of origin, liquid of an andesitic or basaltic nature will come to honeycomb the rock. The content of gases is presumably sufficiently high to reduce the viscosity. The liquid will then become able to transmit hydrostatic pressures, and, although comprising only a part of the rock mass, will constitute a continuous column of considerable height. Then becomes possible the second stage, the draining upward and the convergence of the fluid rock. Gravity is the ultimate cause, as in the downward movement of waters, but here the fluid, being lighter than its surroundings, tends to move upward. An explanation of this draining process has been given by Lane.[1] In a gas-saturated rock an excess of gas, or liquid and gas, would have the power of opening fissures at any depth in the zone of flow without the necessity for the existence of any tensile stress in the walls, or competence in the walls to maintain an open cavity. All that is necessary is that the excess pressure in the rising wedge of gas should be stronger than the cohesion of the rock. The fissure becomes filled with gas and fluid of lesser density than the solid rock of the walls. Consequently, the pressure transmitted from below is greater than the resisting pressure in the walls. This insinuating power, owing to the hydrostatic head due to the lesser gravity of the wedge, becomes greater the higher the wedge rises above the source, until near the surface the action may become violent and rapid. Daly also has outlined a theory of mechanism for the injection of abyssal wedges of magma into the upper crust.[2]

[1] "Geologic Activity of the Earth's Originally Absorbed Gases," *Geological Society of America Bull.*, V (1894), 259–80.

[2] *Am. Jour. Sci.*, XXII (1906), 195–216; *Igneous Rocks and Their Origin* (1914), chap. ix.

His theory is constructed however for action within the roof of a magmatic substratum. Daly postulates a zone of tension, but the mechanism suggested by Lane does not require this and would seem to apply better to the region of generation of magmas, for there the cubic compression is enormous, it is not a zone which has been subjected to cooling, and therefore it is difficult to conceive of the cause of a system of tensile stresses within the asthenosphere.

We are now prepared to draw a closer analogy between the physical conditions of the asthenosphere and those of a glacier, noting the likenesses and also the unlikenesses. In the summer, in the case of Alpine glaciers, heat is supplied to the surface of the glacier until it is warmed to the melting-point, and part of the ice absorbs the amount required by the change of state and passes into water. This trickles along the surface until a fissure is met and the water sinks by force of gravity toward the bottom. Near the snout of the glacier the temperature of this deeper part may thus be more dependent upon this convection than upon direct conduction through the ice. The winter freezing tends to chill the deeper ice and slow its motion, but during the summer the descending water tends to raise the temperature toward the freezing-point. In parts of the glacier where the heating is more effective than the cooling, the waters drill channels and gather at the base of the glacier into streams, reaching finally the outer world. The descent, and gathering, and englacial flow of glacial waters is analogous to the rise and convergence of streamlets of molten rock.

In order to account for lateral mass movement within the asthenosphere, an imperfection of isostasy between continental interior and ocean basin, giving an isostatic gradient or slope toward the continental interior, seems a necessary postulate. Such a gradient is so low that it has not yet been sifted from those irregularities of mass which are owing to the strength of the outer crust, or, it may be, obscured by great compressive bowings of the crust. This isostatic gradient, the slope required to generate movement within the asthenosphere, is far lower than that of the surface of a continental ice sheet. The failures of the analogy between the asthenospheric and glacial states are then as instructive as the agreements. The glacier is thin and broad. Friction on the bot-

tom is excessive and the motion requires more internal work. Much of its mass is permanently well below the freezing temperature. These are the factors which determine the steepness of the surface gradient. The asthenosphere by contrast should be deep and the differential motions within it necessary to satisfy isostasy would be correspondingly small. The temperature through a wide zone should be that of fusion under the hydrostatic pressures prevailing. The solid rock should be sodden with occluded gases, giving mobility to the growing fluid and ready to play their part in assisting recrystallization. Such a physical condition, as long as there is a continuous solid, would exhibit perfect elasticity and high rigidity during the passage of transverse vibrations, yet would slowly yield to prolonged shearing stresses, even though these were very small in amount.

RELATIONS OF IGNEOUS ACTIVITY TO ASTHENOSPHERE AND LITHOSPHERE

The argument has led to the view that the asthenosphere is a region where the temperature curve becomes tangent to the fusion curve, but that a condition of solidity is maintained by the recurrent elimination of that material which becomes molten. The importance of such a process, maintaining the solidity of the earth, has been dwelt upon by Chamberlin, especially as accounting for the overwhelming igneous activity of Archeozoic time. In lessened measure it applies to all later times as well.

Becker has held that the bottom of the zone of isostatic compensation is the depth at which the temperature curve approached nearest to the fusion curve, and he was the first to connect in this way the geodetic evidence with a temperature relation.[1] But Becker does not conceive of actual, permanent contact of the two curves as occurring, and took the depth of nearest approach to fusion as 122 km. This follows from Hayford's hypothesis of a uniform distribution for isostatic compensation, but in the present work there has been found reason for believing that compensation fades out through a greater depth; the strength, as measured by

[1] "Age of a Cooling Globe in Which the Initial Temperature Increases Directly as the Distance from the Surface," *Science*, XXVII (1908), 227–33, 392; "The Age of the Earth," *Smithsonian Miscellaneous Collections*, LVI, No. 6 (1910), 1–28.

the existence of stress-difference, through a depth greater still. The beginning of permanent contact of the two curves, if this is the cause of the disappearance of strength, should be as much as 300 km. deep and extend through some hundreds of kilometers.

A rectilinear projection downward of the temperature gradient observed at the surface would reach the fusion temperature of rocks at a depth of about 50 km. There must be consequently a marked curvature of the temperature gradient if the temperature and fusion curves do not meet short of 300 km. This curvature implies that near the surface there is either a greater quantity of heat flowing outward by conduction or that the conductivity of rock near the surface is very greatly decreased. But such a very great decrease in conductivity making for a higher temperature gradient finds no supporting evidence. On the other hand, a greater outward flow by conduction of heat near the surface may be due to the continued generation of heat by radioactivity to a greater degree than below; or also to a rise of magmas from the asthenosphere. Magmas which never reach the surface would bring heat by a convective process directly into the outer crust. From there the heat, slowly diffused upward by conduction, would increase the temperature gradient in the outermost part of the lithosphere. It is this factor especially which the argument of the present chapter emphasizes.

It is only within the present generation that general recognition has been given to the intrusive nature of the abyssal igneous rocks. They are now generally regarded as risen from the depths. Their action has been to break through and engulf the foundations of the ancient crust.

This process of batholithic invasion seems to be recurrent and widespread, though rise into the outer crust is restricted to the crises of diastrophism and usually reaches levels exposed to erosion only along the lines of mountain systems. The stores of heat brought up from the greater depths would be held in the crust, especially in its deeper parts, for geologic ages, blurring out in the course of time by conduction and creating a false appearance of heat lingering from an initial molten state, a resemblance increased by the added veil of new heat of radio-active origin mantling the ancient stores.

The temperature gradient under this view should naturally vary widely from place to place and from time to time. Igneous activity is the effective means by which heat is brought up from depths which on account of the slowness of conduction would be otherwise thermally isolated from the outer crust. Offset against this, cooling by conduction advances downward from the surface, dissipating not only the heat of local radio-active origin but that excess rising from the depths. The heat of the crust is not then a continually ebbing residuum from a primal molten state, but represents rather an oscillating ebb and flow, one of the balances of nature maintained through geologic time.

If this view be true—that the invasive igneous rocks have been an important factor in determining the amount and distribution of heat in the crust—it is doubtful if any sound arguments can be derived from the study of the present gradients as to the initial temperature. This conclusion is similar to the change of viewpoint in other lines of geology. It was once thought that the composition of the present atmosphere and the character of present climates were steps in a simple and continuous series of changes passing from primal conditions to a future in which the water would be absorbed into the earth and its surface transformed into a frozen desert. Now, however, it is generally recognized that since the earliest known times the surface conditions have been in a state of oscillating equilibrium. The argument of this section leads toward the view that this is true for the physical conditions within lithosphere and asthenosphere also.

[*To be concluded*]

THE BURIED ROCK SURFACE AND PRE-GLACIAL RIVER VALLEYS OF MINNEAPOLIS AND VICINITY

E. K. SOPER
University of Minnesota

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I. INTRODUCTION AND SUMMARY

The rock formations underlying the mantle of glacial drift which covers the region around Minneapolis have been studied and described in detail by N. H. Winchell,[1] Warren Upham,[2] C. W. Hall,[3] F. W. Sardeson,[4] C. P. Berkey,[5] and others. The character of the surface of this buried rock floor has been known only in a general way as revealed by local post-glacial erosion and, in part, by the present surface topography. The definite relations of this buried topography to the present one; the undulations of its surface; and the depth and courses of the buried river channels have been known only at scattered points where data have become available through occasional well borings or other artificial or natural excavations. In the investigations herein described an attempt has been made: (1) to show in detail, so far as available data will justify, the exact nature of the surface of the bed rock; (2) to determine the depth

[1] *Final Report Geol. and Nat. Hist. Survey of Minn.*

[2] *Ibid.*

[3] "Geology and Underground Waters of Southern Minnesota": Water Supply *Paper, U.S. Geol. Survey No. 256.*

[4] "Galena Series," *Bull. Geol. Society of America*, XVIII (1907), 179–94.

[5] "Paleography of Saint Peter Time," *ibid.*, XVII (1906), 229–50.

of the drift at all points within the city of Minneapolis; and (3) to trace the courses of the glacial and pre-glacial river channels, now filled with glacial débris, throughout the city. The results of these investigations are shown on the topographic map on which surface and solid rock is contoured and in the series of geologic structure sections which accompany this paper and to which frequent reference will be made in the following discussion.

The results show the following essential facts. In pre-glacial time the region around Minneapolis was dissected by a large river (referred to tentatively as the pre-glacial Mississippi) and its tributaries, which cut deep valleys into the rock. At a later time, during the Glacial period, the surface of the rock included in the areas between these old stream valleys was also deeply eroded by ice gouging and planation. While the effect of this glacial erosion was to produce a generally flat rock surface above the valleys, there were numerous irregularities of surface developed, many of which have never before been accurately delineated. The ancient valleys became filled with glacial débris which was brought down by the ice and left behind as a thick mantle covering the entire region when the ice sheet receded to the north. Not only were the old valleys choked up with glacial drift, but the entire rock surface was buried. The depth of this drift has been determined at as many points in the city as possible and has been found to vary from 0 to 250 feet. It is thickest in the old buried valleys and thinnest where post-glacial erosion has been active. The depth, width, and courses of the old buried valleys have been determined and are shown in contour on the accompanying map which gives a picture of the buried rock surface throughout the city. In constructing the map (Fig. 1), the surface topography has been superimposed upon this buried topography so that the depth to bedrock at any point may be readily determined from the map by taking the difference between the elevations shown by the surface contours (solid lines) and the bedrock contours (broken lines). The large depression which is the site of Powderhorn Lake was found to be due to the head of a small buried valley in this locality. Five secondary valleys which were tributary to the main buried river valley have been traced. These are shown on the map. Finally, it has been shown that the position of these buried

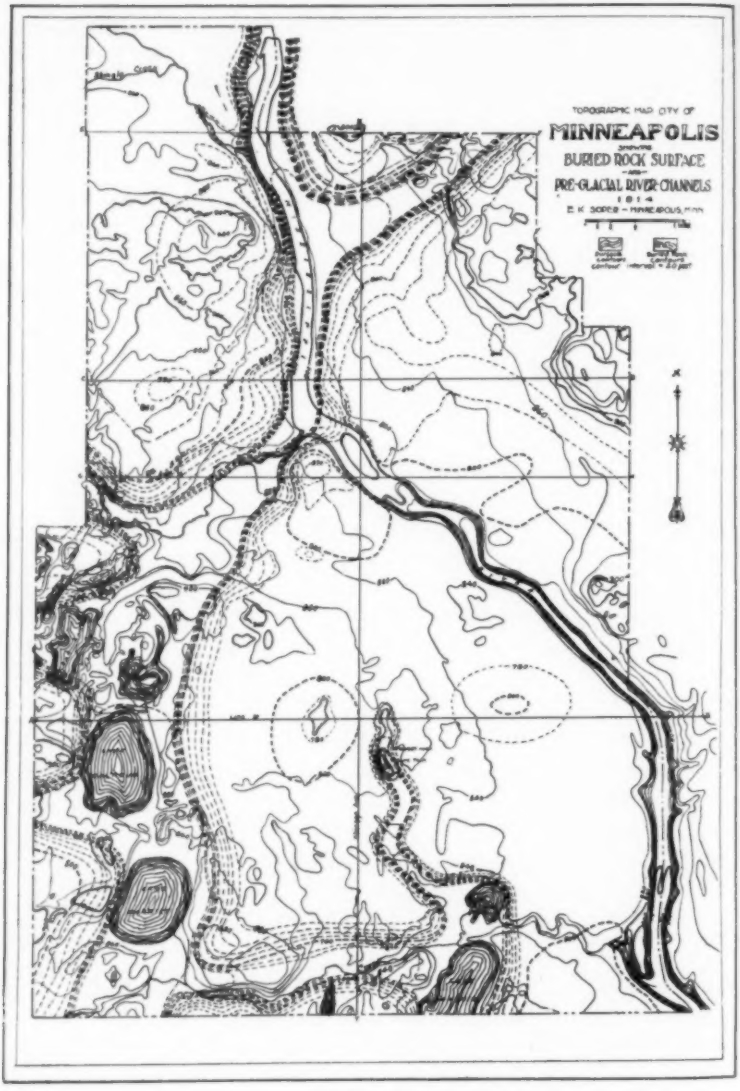
TOPOGRAPHIC MAP CITY OF
MINNEAPOLIS
SHOWING
BURIED ROCK SURFACE
AND
PRE-GLACIAL RIVER CHANNELS
1914
E. K. SOPER - MINNEAPOLIS, MINN.

FIG. 1

valleys should be a controlling factor in the choice of types of foundations for large buildings or other structures erected within the zone.

The data used in constructing the map and sections have been obtained from many sources, but chiefly from records of excavations for sewers and water mains and from hundreds of well records supplied by the well drillers operating in the vicinity of Minneapolis. The surface topography is partly taken from the topographic maps of the United States Geological Survey covering the region around Minneapolis and St. Paul, and partly compiled from elevations supplied by the city engineer, and from field work done in the spring of 1914.

The writer desires to express his thanks to all those who have given assistance in gathering data for this paper. Special acknowledgment is made to City Engineer F. W. Cappelen, Dr. F. W. Sardeson, and to the engineers in the Sewer and Water departments of the City Engineer's office; to Mr. A. C. Godward, engineer for the Minneapolis Board of Park Commissioners; Mr. James G. Houghton, city building inspector; Mr. J. F. McCarthy, Mr. E. F. McCarthy, the S. Swenson Artesian Well Co., and Mr. Max Renner.

II. ROCK FORMATIONS BENEATH THE DRIFT

The rock formations existing in the vicinity of Minneapolis, with the relative position, thickness, and characteristics of each, are shown in Table I.

Only four of the formations given in the table may constitute the rock surface immediately beneath the glacial drift in Minneapolis. These are (1) the Decorah shale; (2) Platteville limestone; (3) St. Peter sandstone; and (4) Oneota dolomite. The existence of the latter in direct contact with the drift has been determined at only two or three localities within the city limits, where deep wells have penetrated the drift. The localities have been found to lie over the deepest part of the main buried river channel, as shown by the deep well records. Therefore it seems probable that the pre-glacial Mississippi had cut entirely through the overlying formations and flowed upon a rock bed of Oneota dolomite from a point near the mouth of Bassett Creek in North Minneapolis southward to the city limits and beyond.

All four of the formations, each of which constitutes the first rock beneath the drift at different places, originally extended over the entire area of the city and far beyond. Before pre-glacial valleys were carved the rock now topmost was continuous over all others. The amount of this surface erosion has been greatest along

TABLE I

GENERAL GEOLOGIC SECTION FOR MINNEAPOLIS

Era	System and Series	Formation	Character of Strata	Approximate Maximum Thickness
Cenozoic	Pleistocene	Glacial drift	Bowlder clay, sand, gravel clay, alluvium, etc.	175 feet
Paleozoic	Ordovician	Decorah shale	Green limey shale	35
		Platteville limestone	Blue to gray thin-bedded limestone	25–35
		St. Peter sandstone	White or yellow fine-grained sandstone with some shale	175
		Shakopee dolomite	Yellow, buff, or pink dolomitic limestone	70
		New Richmond sandstone	White sandstone	40
		Oneota dolomite	Buff to reddish dolomite	100
	Cambrian	Jordan sandstone	Coarse-grained white sandstone	80–100
		St. Lawrence formation	Dolomite, shale, and a little sandstone	200 (?)
		Dresback sandstone	Fine-grained white sandstone, shaley toward base	250
Proterozoic	Algonkian	Red Clastic Series	Red sandstone and shale and some volcanic rocks	0–?
Archean	Archean	Granite	Granite and gneiss	?

the courses of the old abandoned river channel and its tributaries. These ancient streams had cut entirely through the Decorah shale, Platteville limestone, and well into the St. Peter sandstone throughout their courses, and, as explained above, the trunk channel was so deeply eroded as to expose the Oneota in its bed along the lower portion of its course through the city.

This erosion was not confined to the pre-glacial streams but was, in a smaller degree, effective over the entire area. The Decorah shale, which is a persistent formation in Goodhue, Olmsted, Rice and other counties in southern Minnesota, has been almost completely eroded from the area included in the map. Only small isolated areas remain, capping some of the higher points in the south central and southeastern parts of the city and especially along the banks of the present channel of the Mississippi from the University campus southward to Minnehaha Creek.

Not only has pre-glacial and glacial erosion removed most of the shale from the area, but the limestone is also more or less eroded at all points within the city and completely removed from much of the northern part of Minneapolis, especially the northwestern portion along the area drained by Shingle Creek. Nevertheless, the Platteville limestone is the principal rock formation immediately below the drift and covers by far the greater part of the area within Minneapolis. This limestone is absent only along the stream channels, in the northwestern part of the area, and in a few scattered areas in north Minneapolis where erosion has reached the underlying sandstone. The distribution and relations of the formations within the city of Minneapolis may be seen on the map and sections accompanying this paper.

III. THE SURFACE OF THE BURIED ROCK

It will be seen from an inspection of the sections (Figs. 2 and 3) that the rock surface beneath the drift is generally nearly flat. The only notable features of relief are found along the old river channels which have been eroded to an average depth of 200 to 250 feet below the rock surface of the intervening upland plain. There are, however, numerous minor irregularities of surface due to the unequal erosion by the ice. The dip of the rocks is so flat as not to be noticeable to the eye except along the north edge of the city, where there is a low dip. Owing to the scarcity of outcrops this dip is only evident when the elevations of the top of the rock are compared to those in the central and southern part of the city. The top of the limestone along the south edge of Minneapolis stands at an elevation of about 790 to 800 feet above sea-level,

depending upon the amount of erosion which has occurred. In the extreme north edge of the city, in an old quarry about 3,400 feet east of the river, the top of the limestone is about 900 feet above the sea. About a quarter of a mile east of this quarry, underlying the Columbia Heights district, the limestone surface reaches its maximum elevation for the area, about 920 feet. This shows a rise of from 100 to 130 feet in a distance of about 8½ miles. Most of the rise, however, is in the northernmost two miles, where the

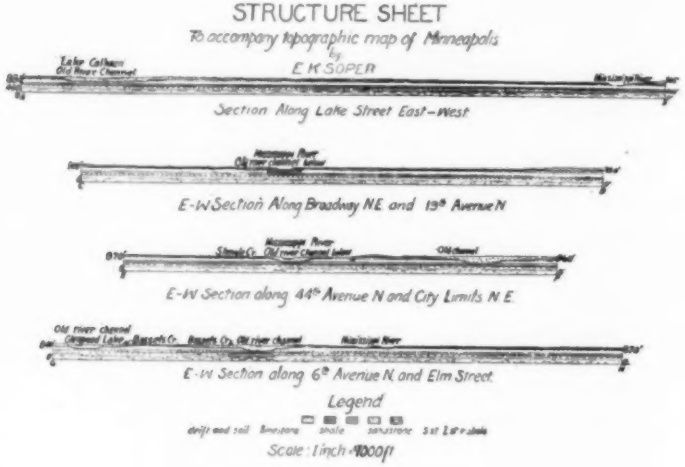
STRUCTURE SHEET
To accompany topographic map of Minneapolis
by
E. K. SOPER
Section Along Lake Street East–West
E-W Section Along Broadway NE and 13th Avenue N
E-W Section along 44th Avenue N and City Limits N.E.
E-W Section along 6th Avenue N. and Elm Street
Legend
drift and soil
limestone
shale
sandstone
Scale: 1 inch = 1000 ft

FIG. 2

dip steepens appreciably. The dip in the south half of the city is practically negligible. The elevation of the sandstone is correspondingly higher in the northern portion of the area, although the rise is not quite as great as in the case of the limestone, owing to the fact that the limestone is several feet thicker in the Columbia Heights district than in the south central portion of the city, where erosion has removed the upper layers. There are several minor undulations in the strata between the north and south city limits, but these are so slight as to be unimportant.

The generally flat surface of this limestone plain, back from the buried river gorges, is broken here and there by small hillocks.

These may be capped by a few feet of Decorah shale, which represent remnants of erosion. There are also several depressions or basins, probably gouged out by ice scouring, which, in several instances, are deep enough to extend through the limestone into the underlying sandstone. Another explanation for these basins is that they represent sink holes caused by the caving of the limestone over solution cavities produced in the underlying strata by ground waters. There is one cave about half a block long at Second Avenue South, between Fourth and Fifth streets, in the downtown district. This has not resulted in any noticeable surface depression, but in order to avoid such a possibility in the future, which might prove disastrous to buildings in the vicinity, it has been thoroughly walled up with concrete supports. This cave was due to mechanical action rather than to solution. The cavity occurs at the contact between the limestone and sandstone, but lies entirely within the sandstone. It is clearly due to the action of the ground water which has washed away the somewhat loosely consolidated sand at the contact, where the ground water tends to accumulate beneath an impervious bed of shale which separates the sandstone from the overlying limestone. This locality is only a few blocks from the river gorge along the sides of which the ground water comes to the surface. The production of open cavities or channels in the sandstone by circulating ground

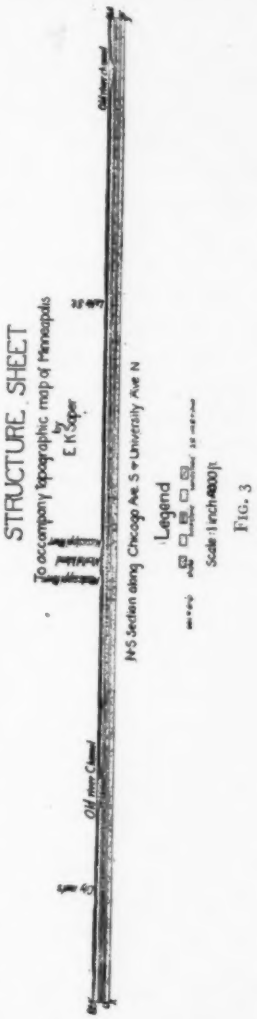

FIG. 3

water was also observed in driving the tunnels through the sand rock on the University campus for the new heating plant. In the course of this work a fissure in the sandstone was encountered through which a large stream of water issued and temporarily flooded the tunnel. With this indisputable evidence of the occurrence of cavities in the rock below the surface, it is entirely possible that certain of the depressions in the limestone surface, as shown by excavations through the drift, may be due to the slumping of the limestone over similar cavities.

IV. RELATIONS OF THE BURIED ROCK SURFACE TO THE EXISTING DRIFT SURFACE

The major features of the present topography of the surface of the ground throughout the city of Minneapolis are in general independent of the underlying rock, and this is what one might expect in a region so heavily glaciated. There are, however, certain relations between the present surface and the buried topography which are suggestive and which, when understood, can be used to advantage in interpreting the character of the buried surface. The localities where there is an evident relation between the two surfaces are along the buried river channels. These old stream courses, which have been filled with débris to a depth of 200 to 250 feet, may still be roughly traced across the city by a series of depressions or low areas in the drift, some of which are now the sites of lakes. These depressions do not form a continuous and unbroken course in the present drift, for occasionally the independence of the glacial topography is manifested by abrupt morainal hills extending across the chain of depressed areas. An inspection of the accompanying map will show that nearly every natural lake within the city lies in the course of some buried stream. Not only are these pre-glacial channels indicated by a chain of lakes and shallow basins, but the general depression of the drift over the old stream beds has, at several places, resulted in determining the direction of the present surface drainage. Bassett's Creek and Minnehaha Creek both flow, for at least parts of their courses, directly over much larger drift-filled channels (see Fig. 1). Even the present channel of the Mississippi River, north of the mouth of Basset's Creek, to the city limits

and beyond, is superimposed upon the buried gorge of the pre-glacial river. Over the greater portion of the city, however, there is no suggestion in the surface topography of the irregularities in the buried rock surface. The topography of the existing surface is typically glacial in its character, and, although the maximum relief is not as great as in the case of the underlying rock surface, yet the general relief is seen to be stronger than that of the underlying rock surface, if we do not consider the old buried gorges. The lowest point on the surface within the city is at the bottom of the Mississippi River just below the mouth of Minnehaha Creek, in the extreme southeastern corner of the city, where the elevation is about 690 feet above sea-level. The highest points are along the ridge of morainal hills in the northeastern part of the city from Columbia Heights to a point about one-fourth mile southeast of Hillside Cemetery, where there are several points which attain an altitude of 965 feet, or thereabouts. This gives a maximum surface relief of about 275 feet. The lowest point accurately known on the surface of the bedrock was determined from the record of a well at the Bath House on the north shore of Lake Calhoun where the surface of the rock was encountered at an elevation of 619 feet. The highest point attained by the rock is in the Columbia Heights district where the top of the shale stands at about 930 feet. This gives a maximum relief of 311 feet for the rock surface, which is 36 feet greater than the relief of the existing drift surface.

There are probably two factors which are mainly responsible for the superimposed drainage system of this region and the general occurrence of low areas along the courses of the buried river valleys. Perhaps the most important one is the great depth to which these old gorges had been eroded. During the process of filling these valleys with glacial débris it may be that the amount of material available was not always quite equal to the amount required to fill them level with the tops of the banks at all points. This would result in slight depressions at intervals along the courses of the former streams. Another important factor is probably the settling of this newly deposited and loosely consolidated material. This settling would result from the saturation of the mass by ground water or from the gradual packing of the filling under its own

weight. In either case the shrinkage in a mass 200 to 250 feet thick might be sufficient to produce low undrained areas on the surface. Where these areas of depression are shallow and unimportant, settling could account for their existence. Where the surface basins are large and deep, as in the areas occupied by Lake Calhoun, Lake Harriet, Lake Nokomis, and that portion of the Mississippi River gorge north of the mouth of Bassett's Creek, their origin may be due to the incomplete filling of the pre-glacial gorge.

Because of the relatively even erosion over the flat areas between the pre-glacial valleys, there are few prominent peaks or high points of rock beneath the drift, and hence few outcrops are to be found in the city. The principal rock exposures are found along the post-glacial valley of the Mississippi from St. Anthony Falls to Fort Snelling and along the gorge of Minnehaha Creek, which is also post-glacial in age, from its mouth to Minnehaha Falls. Also at a few localities in North Minneapolis, where the drift is relatively thin, post-glacial erosion or artificial excavation has exposed the bed rock. There are also a few exposures near water-level on Nicollet Island and northward along the river to the mouth of Bassett's Creek.

V. THE BURIED RIVER CHANNELS

The history of the buried river channels of Minneapolis and vicinity, with an account of the origin and recession of St. Anthony Falls, has been worked out by N. H. Winchell,[1] U. S. Grant,[2] F. W. Sardeson,[3] and others. The general course of the main pre-glacial channel of the river across Minneapolis has been known for many years, but there has been no attempt to determine accurately the boundaries of the river and its tributaries and to construct a topographic map of the old rock surface beneath the drift that would

[1] *Geol. Survey of Minn., Fifth Annual Report* (1877), 175; also *Final Report*, II (1888), 313.

[2] "An Account of a Deserted Gorge of the Mississippi River near Minnehaha Falls," *American Geologist*, V (1890), 1.

[3] "The Beginning and Recession of St. Anthony Falls," *Bull. Geol. Society of America*, XIX (1908), 29–52.

show the depth and width of these buried gorges throughout their courses. The lack of data in former years made it impossible to construct such a map with any degree of accuracy. During the last decade, however, much additional data have been made available, especially at critical localities, through the sinking of wells and the construction of sewer tunnels and ditches and other excavations, so that it has become possible to show in some detail the character of the old drainage system. As will be seen from a study of the topographic map, the limestone which immediately underlies the drift over most of the city has been cut into six segments by pre-glacial stream erosion, and had St. Anthony Falls cut its way about one mile farther up the Mississippi to the mouth of Bassett's Creek before its progress was arrested by artificial means,[1] there would have been seven separate areas of limestone instead of six.

The present river follows the ancient course, which it has partially re-excavated, from a point beyond the north city limits to the mouth of Bassett's Creek in the north central part of the city. At this point the river leaves its pre-glacial channel, turns to the east, and flows southeastward through a young narrow gorge to Fort Snelling, at which point it is joined by the Minnesota River which flows into it from the south. As the map shows, the ancient river valley continued southward from the point where Bassett's Creek now has its outlet and flowed in a broad valley with gradually sloping banks southward across the region now occupied by the chain of lakes which comprises Lake of the Isles, Lake Calhoun, and Lake Harriet. At Bryn Mawr Meadows a tributary stream joined the river from the northwest, and its buried channel is now the course of Bassett's Creek between Glenwood Lake and Bryn Mawr Meadows. Glenwood Lake also lies within this tributary valley. A second tributary stream joined the main river from the west, just opposite Lake of the Isles. The presence of a buried valley east of this point is suggested by a series of basin-like depressions, one of which is the site of Cedar Lake. The existence of a third tributary valley opposite the south end of Lake Calhoun is indicated by well

[1] A protecting apron of timber was built over the rock at the point of fall to protect the rock from further erosion, which threatened to destroy the falls.

records. This valley also joins the main valley from the west, but there is little evidence of its existence in the present surface topography, which consists of low hills of drift and completely obliterates any underlying channel.

The pre-glacial river flowed almost due south from Lake Harriet, but no attempt has been made to outline its exact course beyond the city limits. In a general way this course can be traced for about five miles south of Minneapolis through Grass Lake, Wood Lake, and numerous small lakes and ponds to its junction with the Minnesota River valley somewhere east of the village of Bloomington.

An important tributary valley had its confluence with the main valley just south of Lake Harriet, at which point it entered from the east. Evidence of this buried valley may be seen in the present topography, especially along Minnehaha Creek, which follows the course of the ancient channel, between Lake Harriet and Lake Nokomis. At the latter point the old valley turns sharply to the south.

At Powderhorn Lake a small valley heads and flows southward to Rice Lake where it joins the tributary just described. This region around Powderhorn Lake is particularly interesting in this connection, for it has usually been considered to be a large Kettle hole.

North of the mouth of Bassett's Creek the Mississippi River flows over the same course that was formerly occupied in pre-glacial time. The old channel has been almost filled with sediments and drift so that the present stream bed, which consists of glacial débris and recent clays and silts, lies in this region at an elevation of at least 160 feet above the bottom of the buried valley. At a point about half way between the mouth of Bassett's Creek and the north city limits, along the river between Twenty-sixth Avenue North and Thirty-first Avenue North, a large tributary valley joined the pre-glacial river. This buried valley follows a course indicated in the present topography by a low swampy zone extending northeastward from the river through Sandy Lake (now dry) across Columbia Park, and beyond the city to Silver Lake. The exact course of the valley north of Silver Lake has not been determined.

The character of the material filling the ancient gorge in the north part of the city between the mouth of Bassett's Creek and the north city limits is strikingly different from that which constitutes the valley filling at all other points in the course of the main gorge and its tributaries. Along this particular portion of the river course the old valley is filled with fine laminated clays and silts instead of the heterogeneous drift which characterizes the filling at other places. The upper portions of these clay beds have been used for many years in the manufacture of brick and tile. These fine silts seem to extend from the bed of the present river (which at this place flows directly over the pre-glacial gorge) down to the bottom of the buried valley, a distance of about 160 feet. These silts also are found back from the main channel, where they cover many acres along the flats on both sides of the river. Their thickness is much less, however, away from the river.

This sudden change in the nature of the material filling the old valley, the suggestive laminated structure of the upper portion of the deposit, the variation in thickness of the material, and its occurrence back from the river channel, along the flood plain of the river, are all evidences which point to an original deposition in a lake. The change from gravel and drift to clay and silt occurs immediately above the point where the present river abandons the pre-glacial course. It seems probable that at a late stage in the recession of the ice a moraine or dam was formed across the old gorge near the mouth of Bassett's Creek. This resulted in the formation of a temporary lake which filled the river valley, overflowed the banks, and spread over the low land which flanked the valley on both sides. In this lake the fine clays and silts could be deposited. Eventually an outlet was formed for the impounded waters at the point where the present river diverges from the filled gorge. After the outlet was once established the erosion of the new post-glacial gorge was started. But deposition of fine sediments could continue for a long time after the new gorge was started, because the lake would still remain and the current would continue to be slackened to the point where clays and slits could accumulate. The boundaries of this lake have not been traced north of the city limits, but it is probable that it extended a considerable distance beyond them.

VI. THE DRIFT

The glacial drift which was spread over the land upon the retreat of the ice, and which covers the rock surface throughout the city, is variable in its composition. It ranges from typical bowlder clay, which consists of a heterogeneous mixture of clay, sand, gravel, and bowlders, to sorted deposits of fine sand or clayey loam. Most of the drift consists of a mixture of clay, sand, and bowlders of all sizes. Where the sand in the drift is free from clay and gravel, it may have been deposited in two ways: either it represents a wind-blown deposit and originated as sand dunes, or it is a flood-plain deposit, washed up by glacial streams and sorted by water action. The only place in the city where clay occurs in large quantity free from pebbles or bowlders is along the banks of the Mississippi River north of the mouth of Shingle Creek. These clay beds probably represent lake and river deposits rather than accumulations due to glacial action. They are indirectly the product of glacial action, since they were deposited during the Glacial stage of the river.

Since the surface of the drift constitutes the surface of the land, except where it has been covered by post-glacial deposits of alluvium, sand, or *loess* of slight thickness, it controls the topography of the region. Therefore, what has been said concerning the characteristics of the topography of the city will apply also to the drift.

The thickness of the drift is variable and ranges from 0 to 250 feet. As shown by the scarcity of outcrops (see map) the drift is present nearly everywhere except along the river bluffs. The localities where it is thinnest are those which border the rock outcrops. In north Minneapolis there is an area bounded by Central Avenue on the south and east, Twenty-third Avenue on the north, and Main Street on the west, where the drift is thin, ranging from 0 to 15 or 20 feet. Throughout the greater part of South Minneapolis, in the large area bounded by the Mississippi River on the east and north, the pre-glacial Mississippi River valley on the west, and Minnehaha Creek on the south, the thickness of the drift is very uniform. The average depth over most of this area is from 30 to 50 feet, the greatest thickness being in the southern portion. Both the surface of the drift and the surface of the underlying limestone are generally flat over this region. The only noteworthy breaks in

the otherwise almost flat surface are at Powderhorn Lake, where a small buried valley heads, and at Lowry Hill and the region around Loring Park, where a belt of low morainal hills extends from the west across the edge of the area and breaks the monotony of the surface.

In some portions of the city, especially in Southeast Minneapolis, in the vicinity of the University and northeastward to the city limits, there is a bed of peat several feet thick which occurs only a few feet below the surface. This peat bed rests upon drift largely composed of sand, clay, or gravel, or a mixture of the three, and is covered by recent deposits of sand or soil which are of post-glacial or glacial origin.

In East Minneapolis, near Tower Hill in the Prospect Park district, deposits of *loess*, a formation of loamy material of aeolian origin, occur at the surface. Therefore, it is apparent that while the retreating ice left a mantle of drift over the entire region, the surface of this drift has been slightly modified by post-glacial deposition. These deposits of sand, silt, and other material were chiefly laid down as flood-plain deposits in the streams and marshes fed by waters from the melting ice. A small amount of post-glacial erosion has also operated to modify the original drift surface, but this has been unimportant except along the channel of the river and its tributaries.

The origin, composition, and character of different drift deposits and their modification by post-glacial agencies have all been described in detail in reports by N. H. Winchell,[1] Warren Upham,[2] F. W. Sardeson,[3] and F. F. Grout and E. K. Soper.[4]

VII. INFLUENCE OF BURIED RIVER CHANNELS UPON BUILDING CONSTRUCTION

In those portions of the city which are underlain by the buried valleys and in other parts of the city where the drift is deep, it is impracticable to carry excavations to bedrock for building foundations and cellars. For the ordinary small store or residence, or for

[1] *Final Report Minn. Geol. and Nat. Hist. Survey.* [2] *Ibid.*

[3] "Geology of the Twin Cities," *Minneapolis and St. Paul Folio, U.S. Geol. Survey.* Now in press (1914).

[4] Grout and Soper, "The Clays and Shales of Minnesota," *Bull. No. 11, Minnesota Geological Survey*, chapter on Hennepin County (1914).

light buildings of any description, it is not necessary to have a perfectly rigid foundation. The drift throughout Minneapolis is usually fairly solid at moderate depths below the surface and is quite safe for foundations for buildings of average size. But in the regions along the courses of the buried river valleys, where the drift may be from 100 to 250 feet deep, areas of loose sand which often has a tendency to run are frequently encountered. In such localities it has been found necessary to use piling in constructing foundations for large buildings.

The main business section of the city lies back from the buried valley and in the downtown district the average distance to bedrock is only 25 to 30 feet. The rock in this locality consists of limestone or hard shale and can easily be reached if necessary in building operations. For this reason it has been necessary to use piling in the construction of only a few buildings in the city. Should the zone of large buildings north of Hennepin Avenue be extended in the future a few blocks farther west beyond the Great Northern Railroad tracks, it will be found necessary to use piling, caissons, or some type of foundation adapted to soft ground, for many if not all of the large structures. The side of the buried valley in this vicinity drops off rapidly, so that points along the present channel of Bassett's Creek overlie the very bottom of the old valley nearly 200 feet below. Much of the material filling the old gorge in this vicinity is loose sand. At all points in the city, except those along the sides and courses of the buried valleys, the foundations are generally good.

SUMMARIES OF PRE-CAMBRIAN LITERATURE OF NORTH AMERICA FOR 1909, 1910, 1911, AND PART OF 1912

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IV. ONTARIO, NORTHWEST TERRITORIES, AND THE EAST COAST OF HUDSON BAY

Allen[1] states that the Woman River area in the Sudbury Mining Division of Ontario, west of Rush Lake, shows the following succession:

Basic igneous dikes, Mica porphyry } Relative ages not known
Acid igneous rocks—extrusive and intrusive
Iron formation
Basal greenstones

The series has been intensely folded and extensively brecciated. The strike of the iron formation banding is N. 45° E., but is in some places at right angles to this direction where modified by cross-folding. Everywhere the rocks stand practically on edge. The basal greenstones and the iron formation are intruded by a great many basic and acid dikes, the latter showing gradation into an acid volcanic breccia overlying the iron formation.

The various phases of the iron formation are: (1) finely banded cherty carbonates, (2) hematitic, magnetic, pyritic, cherts, (3) black and red jaspilites, (4) a unique amphibole-magnetite rock, in which the petrographic character of the amphibole suggests riebeckite, and (5) iron ore. The cherty iron carbonates are regarded as the source of the other iron-bearing rocks, although the possibility that some of the varieties may be partly original is not denied. The derivation of the secondary varieties is ascribed to the same katamorphic and anamorphic processes by which Van Hise has explained the origin of similar derivatives in the other iron formations.

[1] R. C. Allen, "Iron Formation of Woman River Area," *Eighteenth Annual Report of the Bureau of Mines*, 1909, Ontario.

The distribution of the iron ores bears no relation to the present structural features of the formation nor to its present erosion surface.

The relation of the iron formation to the associated igneous rocks is expressed by the following statements: The iron formation is sedimentary; it conformably overlies the ellipsoidal greenstones, in that no period of subaerial erosion intervened between the deposition of the iron formation and the greenstone; the iron formation and the overlying volcanic breccia and rhyolite porphyry are also conformable. The evidence of the sedimentary origin of the iron formation consists in its banding, the iron carbonate content, and the parallelism of the individual bands to each other and to the basal plane.

The evidence for conformity with the basal greenstone lies in the parallelism of the banding of the iron formation with the plane of contact with the greenstone, the absence of detrital material, and the probable subaqueous origin of the ellipsoidal greenstones. The conformity of the iron formation and the overlying acid igneous rocks appears probable but is not supported by many field observations, owing to the sparsity of exposed contacts. Wherever observed, the contacts are sharp and show no detrital materials.

From field relations of the iron formation to igneous rocks in the Woman River and other areas but mainly from the work of Van Hise and Leith, Allen believes that the genesis of the iron formations was related to igneous rather than sedimentary agencies and processes. He interprets the physical history of this area to be extrusion of basic greenstones upon a submerged floor, rapidly followed by the precipitation of iron formation, contributed from magmatic solutions, without the intervention of clastic sedimentation. The precipitation of the iron formation was followed by the extrusion of acid igneous rocks. The three rock types, however, basal greenstone, iron formations, and acid igneous rock, are perhaps differentiates of one parent-magma.

Adams[1] describes the occurrence of silver and calcite, in wall rock adjacent to a fissure vein of the Cobalt district.

[1] F. D. Adams, "Notes on the Occurrence of the Ore Body at the City of Cobalt Mine," *Jour. Can. Min. Inst.*, XII (1909), 414–17.

Baker[1] states that the pre-Cambrian rocks of the Lower Mattagami basin are pink gneisses of Laurentian age, Post-Middle Huronian diabase dikes, a small outcrop of a banded siderite, quartz conglomerate, and quartzite which may be Upper Huronian.

Baker[2] states that the pre-Cambrian in the Lake Abbitibi area consists of Keewatin schistose, ellipsoidal greenstones, and other metamorphosed eruptive rocks infolded with dolomites, graphitic slates, jaspilites, and fragmental materials, all intruded by Laurentian granites, pegmatites, and granite porphyries. This old complex is intruded by post-Huronian quartz diabase, quartz gabbro, and Camprophyric and aplitic dikes. The diabase is associated with veins showing traces of silver and gold, but no important metal deposits have thus far been found.

Baker[3] states that the pre-Cambrian rocks of the Mattagami River region are Laurentian granites and gneisses intruded by diabase dikes, probably Post Middle Huronian in age, Upper Huronian siderite, sideritic conglomerate, and quartzite.

Bancroft[4] reports the following pre-Cambrian formations from the Keekeek and Kewagama Lakes region, 40 miles northeast of Lake Temiskaming.

Post Huronian	—dikes and stocks of basic intrusives probably Keweenawan
Huronian (?)	—conglomerate, arkose, and graywacke—all schistose
Laurentian	—batholiths and stocks of granite and syenite
Keewatin	—basic and acid schists, iron formation, tuffs, volcanic breccias, rhyolites, basalts, basic and acid extrusives

Burrows[5] states that the South Lorrain silver area is north of the Montreal River and west of Lake Temiskaming, and about 16 miles southeast of the Cobalt district. The pre-Cambrian rocks are Keewatin, basic igneous rocks, and minor acid intrusives, intruded

[1] M. B. Baker, "Iron and Lignite in the Mattagami Basin," *Ont. Bur. Mines, 20th Ann. Rept.*, pp. 214–46, 37 figs.

[2] M. B. Baker, "Lake Abbitibi Area," *Ont. Bur. of Mines, 18th Ann. Rept.*, XVIII (1909), 263–83, pl. 1, 9 figs.

[3] M. B. Baker, "The Iron Ores of the Mattagami River," *Jour. Can. Min. Inst.*, XIV, 299–309.

[4] J. Austin Bancroft, "Report on the Geology and Mineral Resources of Keekeek and Kewagama Lakes Region," *Province of Quebec*, 1911, pp. 160.

[5] "South Lorrain Silver Area," *Ont. Bur. Mines, 18th Ann. Rept. (1908) pt. 2*, pp. 21–31, 1909.

by Laurentian granite and syenite; Huronian quartzite, arkose, conglomerate, slate, and breccia, unconformably above the Keewatin and Laurentian; and post-Huronian diabase intrusives. Quartz and calcite veins bearing native silver, smaltite, and niceolite have been found near contacts between the post-Huronian diabase intrusives, and Keewatin basic igneous rocks.

Burrows[1] reports that the pre-Cambrian succession in the Gowganda and Miller Lake silver area between Sudbury and Cobalt is Keewatin altered basic igneous rocks and acid porphyries, intruded by Laurentian granite, syenite, and gneiss; unconformity; Huronian quartzite, arkose, graywacke, conglomerate, and slate; intrusive into all the preceding, post-Middle-Huronian diabase. The diabase contains silver-bearing veins whose value has not been determined.

Burrows[2] finds that the compact rocks of the Porcupine district 100 miles northwest of the Cobalt district, are all pre-Cambrian, principally Keewatin. Keewatin and Huronian rocks contain irregular gold-bearing veins, in which quartz is the dominant mineral.

Bowen[3] reports that sills and dikes of diabase cut older formations in the Gowganda Lake district. The dikes, some of them 250 feet wide, show no evidence of differentiation. Red spots composed of an intergrowth of quartz and soda-rich plagioclase feldspar occur in the diabase of the sills. At the contact of sills and slate, a granophyre consisting of soda-rich plagioclase, quartz, and accessory minerals including calcite grades into a slate adinole developed by contact metamorphism. The granophyre and adinole are both ascribed to hydrothermal actions, the granophyre being regarded as an extreme phase of the alteration of the slate, a recrystallization from a state of aqueous fusion.

To the reviewer, it seems that the red spots in the sills, and the granophyre, may be due to the same causes, since they are

[1] A. G. Burrows, "The Gowganda and Miller Lakes Silver Area," *Ont. Bur. Mines, 18th Ann. Rept.*, XVIII (1908), Pt. 2, pp. 1–20, 20 figs., 3 maps.

[2] A. G. Burrows, "Porcupine Gold Area," *Quart. Bull. Canadian Min. Inst.*, No. 16, 1911, pp. 59–62.

[3] Norman L. Bowen, "Diabase and Granophyre of the Gowganda Lake District, Ontario," *Jour. Geol.*, XVIII, No. 7 (1910), 658–74.

mineralogically similar and are gradational. Bowen believes that the granophyre may be quite different in origin from the red spots scattered through the sills, despite their similarity. A somewhat similar phenomenon in the Cobalt district, the separation of diabase and aplite, has been ascribed to differentiation by Collins and Hore, Collins showing that the separation depended on the rate of cooling, being entirely absent in diabase, which was rapidly chilled. A magmatic separation of soda-rich aplite and gabbro in the intrusive of Mt. Bohemia, Mich., has been described by F. E. Wright. Here there is no evidence that the aplite was associated with slate.

Bowen[1] presents chemical analyses and descriptions of the diabase and aplite of the Cobalt silver area.

Bowen[2] describes the rocks of the Thunder Bay district as Upper Huronian argillites, gray quartzite and gray slates interstratified, black slate, and iron formation, all unconformable above the Laurentian granite complex, and intruded by Keweenawan diabase dikes and sills. Silver veins are found in steeply dipping fault zones cutting the black slate in an east-west direction. The ore minerals are principally native silver and argentite, associated with calcite, quartz, fluorite, barite, witherite, and other minerals.

Coleman[3] states that the rocks of the Black Sturgeon Lake district southwest of Lake Nipigon are Keewatin green schists intruded by granite gneiss, and slightly disturbed Keweenawan diabase, shale, and sandstone. Hematite veins are found in fault zones of the Keewatin.

Coleman[4] reports that the Alexo nickel deposit near Matheson, northern Ontario, consists of a pyrrhotiferous nickel-bearing rock of unknown extent, grading upward into an ultra basic rock, peridotite altered to serpentine, and resting on footwall of andesite.

[1] N. L. Bowen, "Diabase and Aplite of the Cobalt-Silver Area," *Jour. Can. Min. Inst.*, XII (1909), 517-28.

[2] N. L. Bowen, "Silver in Thunder Bay District," *Ont. Bur. Mines, 20th Ann. Rept.*, pp. 119-32.

[3] A. P. Coleman, "The Black Sturgeon Lake District," *Ont. Bur. Mines, 18th Ann. Rept.*, Pt. 1, pp. 163-179.

[4] A. P. Coleman, "The Alexo Nickel Deposits," *Econ. Geol.*, V, No. 4 (1910), Pt. 4, 372–76.

Coleman interprets the deposit as a marginal magmatic segregation from the peridotite.

Collins[1] finds that the pre-Cambrian rocks in the region between Lake Nipigon west of Clay Lake, a distance of about 220 miles, include Keewatin and Huronian altered acid and basic intrusives, volcanic tuffs, chlorite, and sericite schist with bands of conglomerate, graywacke, phyllite, quartzite, and jaspilite; Laurentian granites, syenites, and diorites intrusive into the Keewatin; post-Keewatin eruptives, gabbro, mica syenite, acid porphyries, and granite; flat-lying Keweenawan impure dolomite and sandstone, and post-Keweenawan diabase intrusive sheets.

Collins[2] reports that the pre-Cambrian rocks between the Pic and Nipigon rivers of the north shore of Lake Superior include Laurentian granite, gneisses, and syenite; Keewatin graphitic gneiss, quartzite, arkose, amygdaloids, and basic and acid schists; Keweenawan red arenaceous dolomite, and diabase intrusives; eruptives of unknown age, viz., syenites, diorite, pegmatite, and diabase.

Collins[3] states that the cobalt silver ores of northern Ontario are associated with diabases. The diabasic magma has acted as a mixture of two rock species, diabase and aplite, whose segregation was controlled by the rate of cooling, no separation having taken place in rapidly cooled rocks. The aplites show differentiation into quartz veins, and still later quartz calcite veins. The diabase contains some chalcopyrite, the aplite still more, while an even higher percentage is contained in the quartz calcite veins. Neither silver nor cobalt were found in the diabases.

Harvie[4] states that the succession of Keewatin, Laurentian, and Huronian rocks of the Oposatica district, about 40 miles east of

[1] W. H. Collins, "A Geological Reconnaissance of the Region Traversed by the National Transcontinental Railway between Lake Nipigon and Clay Lake, Ont.," *Canada Geol. Surv. Branch*, 1909. 67 pp., 2 pls., 1 fig., 2 maps.

[2] W. H. Collins, "Report on the Region Lying North of Lake Superior betwen Pic and Nipigon Rivers, Ont.," *Canada Geol. Survey*, 1909, 24 pp., 1 map.

[3] W. H. Collins, "Quartz Diabases of Nipissing District," *Econ. Geol.*, V, No. 6 (1910), 538–50.

[4] Robert Harvie, "The Telluride Ores at Oposatica," *Jour. Can. Mining Inst.*, 1911, pp. 164–70.

north from the north end of Lake Temiskaming, is similar to that of the Cobalt district. Narrow gold-bearing quartz ankerite veins are found cutting Keewatin and Huronian rocks in the Oposatica district. The gold minerals are petzite and native gold. The petzite occurs in cracks in the quartz and ankerite, and the native gold was deposited after the petzite. Harvie believes that the Oposatica ores will show no increase in values with depth, since the Tellurides in known districts have all been found below the weathered zone.

Hore[1] discusses the stratigraphy and the ores of the Cobalt district. The richest ores have been found in the Huronian conglomerate. Rich shoots have also been mined in the Keewatin greenstone, and in the Keweenawan diabase. The ores are genetically connected with the diabase intrusions.

Hore[2] argues that the cobalt silver ores, and the aplites and diabases with which they are associated are all differentiation products of one magma.

Hore[3] believes that the glacial and glacio-fluvial origin of a portion of the conglomerate-quartzite-shale series of Huronian age at Cobalt and Temagami is clear. The evidence consists in the small amount of stratification, the presence of striated and soled pebbles, the heterogeneity of the conglomerate, and the discordance between the lithologic character of the conglomerate and the basement on which it was deposited. The basal conglomerate may not be of glacial origin since they are like the materials on which they lie. Some of the materials are probably water deposits, into which large erratics were introduced by floating ice. Certain thick bowlder-free deposits of shale and graywacke are probably ordinary water-laid sediments.

Hore[4] states that Keewatin and Huronian rocks similar to those of the Nipissing area outcrop in the Porcupine district. The

[1] R. E. Hore, "Geology of the Cobalt District," *Trans. Am. Inst. Min. Eng.*, XLII (1912), 481–98.

[2] Reginald E. Hore, "Differentiation Products in Quartz Diabase Masses of the Silver Fields of Nipissing, Ontario," *Econ. Geol.*, VI (1911), 51–59.

[3] Reginald E. Hore, "Glacial Origin of Huronian Rocks," *Jour. Geol.*, XVIII, No. 5 (1910), pp. 459–67.

[4] Reginald E. Hore, "On the Nature of Some Porcupine Gold Quartz Deposits," *Jour. Can. Min. Inst.*, XVI. No. 15, pp. 57–70.

Keewatin rocks include acid and basic extrusives, some plutonics jaspilitic iron formation, ferro-dolomites, and schists of unknown origin. The Huronian rocks are chiefly quartzite, graywacke, micaceous schists, and one conglomerate, apparently made up of igneous materials.

Most of the gold ores occur in quartz-bearing veins within metamorphosed igneous rocks of the Keewatin or the Huronian conglomerate. The quartz is full of liquid inclusions and in places is shattered. Gold occurs between quartz veins, included in quartz, in pyrite, and in calcite.

Hore believes that the ores and quartz pyrite gangue were deposited by hot alkaline solutions. He finds that they resemble the ores of the mother-lode closely and that the same hypothesis of origin is applicable to both.

The pre-Cambrian succession in the Nipissing district according to Hore[1] is:

I. Algonkian.
  a) Keweenawan. Igneous intrusions of quartz diabase, quartz gabbro with acid differentiation products. Some olivine diabase and diabase porphyrite dikes. Igneous contact.
  b) Huronian.
    (1) An upper series, probably Middle Huronian. Chiefly feldspathic quartzite with some conglomerate. Barlow's Lorain series. Slight unconformity.
    (2) A lower series, probably Lower Huronian, graywacke, shale, conglomerate, and feldspathic quartzite. Miller's Cobalt series. Great unconformity.

II. Archaean.
  a) Laurentian. Granites, diorites, syenites, gneisses intrusive into Keewatin.
  b) Keewatin. Basic and acid extrusives and intrusives, felsite agglomerate, iron formation, slates and green schist.

The silver ores occur principally in narrow vertical or nearly vertical fissure veins cutting Keweenawan diabase, the lower Huronian series, and the Keewatin. The veins strike in various directions, but are uniform in composition, consisting mainly of smaltite, calcite, and silver, deposited in the order in which they

[1] "The Silver Fields of Nipissing, Ontario," *Can. Min. Inst. Quart. Bull.*, No. 17 (1911), pp. 81–105, 3 pls.

are named. The richest veins are in the Huronian. Hore supports the views that the silver-bearing solutions came from the Keweenawan diabase, that the Keewatin greenstone played an important rôle in their deposition, and that the coarse Huronian sediments afforded abundant openings for their reception.

Hore[1] states that the diabase of the Cobalt district occurs in dikes and sills intrusive into Keewatin schists, and Huronian sediments. The minerals of the diabase are pyroxenes, plagioclase, quartz ilmenite, biotite, and olivine. Olivine seems to be confined to small masses of diabase showing no differentiation. Aplitic micropegmatitic intergrowths of sodic feldspar and quartz, sometimes with calcite, occur in the thick sills. Some of the micropegmatite is clearly a differentiate of the diabase *in situ*, occurring in spots and stringers grading into the diabase. Some of it is a later differentiate, filling fissures sometimes with selvage borders.

A minor fissure filling in the diabase consists of quartz, with chlorite and amphibole. Galena, pyrite, and chalcopyrite are frequently present. Calcite sometimes fills interstices between well-formed crystals of calcite.

King[2] describes a lens-shaped deposit of low phosphorus, nontitaniferous magnetite ore, which occurs at the contact between Grenville limestone and Laurentian granite at the Wilbur mine, in Lanark County of eastern Ontario.

Knight[3] presents diagrams showing faults in the Cobalt district. The largest known fault has a displacement of over two hundred feet. It is of the reversed type and strikes parallel to the longer axis of Cobalt Lake.

Kerr[4] presents a petrographic description of the Port Caldwell syenite area on the north shore of Lake Superior about 125 miles east of Port Arthur. The principal rock of the region is nepheline

[1] Reginald E. Hore, "Diabase of the Cobalt District, Ontario," *Jour. Geol.*, XVIII, No. 3 (1910), 270f.

[2] Shirley King, "The Wilbur Iron Mine," *Jour. Can. Min. Inst.*, XII (1909), 582–91.

[3] C. W. Knight, "Recent Underground Development Work at Cobalt," *Jour. Can. Min. Inst.*, XV, No. 18 (1912), pp. 21–23, 1 pl.

[4] H. L. Kerr, "Nepheline Syenites of Port Caldwell," *Ont. Bur. Mines, 19th Ann. Rept.*, Pt. 1, pp. 194–232, 20 figs., 1 map.

syenite with small amounts of hornblende, augite, and magnetite. Several other syenite facies are represented, besides pegmatites, small bodies of gabbro, granite, diorite, and later basic dikes.

Lane[1] presents a detailed description of the records of 17 drill holes, placed along a line normal to the strike of the Keweenawan copper-bearing rocks on Point Mamainse on the east shore of Lake Superior. The copper-bearing series in this cross-section consist of roughly parallel beds of conglomerate, aplite, and porphyrite intruded by felsite. Their strike is about N. 10° W. and their dip is from 20°–50° west or lakeward. The beds are frequently offset by dip faults.

Leith[2] states that the Algonkian, Nastapoka, and Richmond groups, exposed on the east coast of Hudson Bay, consist of slightly metamorphosed chemical and mechanical sediments, interbedded with basic eruptives, with a seaward dip from 50° to 45°. The Richmond group, unconformably, below the Nastapoka, embraces coarse, ill-assorted mechanical sediments interbedded with basic eruptives, generally dipping seaward at steeper angles than the Nastapoka. The Nastapoka contains an iron formation containing greenalite, iron carbonate, and derivative phases closely related to extrusives. The Nastapoka is probably late Algonkian, perhaps Animikie, because of its structural, stratigraphic, and metamorphic similarity to the Animikie of the Lake Superior region. Leith suggests that during the Algonkian, the Archaean protaxis may have separated the Lake Superior geo-syncline of deposition on the south from a similar region of deposition on the north.

Mickle[3] finds that one hundred square miles of the Nipissing region have numerous calcite veins, but that the metal-bearing veins are restricted to ten square miles. On these ten square miles there are about 2,000 calcite veins of which about 1 per cent are metal-bearing, and only 66 veins were productive at the end of 1907. About 24 per cent of the area is underlaid by Huronian

[1] A. C. Lane, "Diamond Drilling at Point Mamainse, Province of Ontario," *Canada Dept. of Mines, Bull. No. 6*, 1911, 59 pp., 1 fig., 1 map, 5 pls.

[2] C. K. Leith, "An Algonkian Basin in Hudson Bay," *Econ. Geol.*, V, No. 3, pp. 227–46.

[3] G. R. Mickle, "The Probable Number of Productive Veins in the Cobalt District," *Jour. Can. Min. Inst.*, XIII (1910), 325–35.

rocks, 42 per cent diabase, and 33 per cent by the Keewatin. The Huronian had 53 productive veins, the Keewatin 6, and the diabase 7. Assuming that these areas have prospected equally well, the chances of finding ore are about 9 times better on the Huronian than on an equal area of Keewatin, and about one-twelfth as good on the diabase as on an equal area of Huronian.

Miller[1] reports that the pre-Cambrian rocks exposed in the district of Patricia, comprising over 146,400 square miles west of James Bay, which has recently been added to the Province of Ontario, include limestones similar to those of the Hastings and Grenville districts, a small volume of Lower-Middle Huronian conglomerates, iron-bearing Animikie, and diabase intrusives, probably Keweenawan.

Miller and Knight[2] state that they have found an unconformity between the Grenville and the Hastings series of eastern Ontario, the Hastings being younger. The Grenville, they maintain, has been deposited on rocks similar to the Keewatin greenstones of Lake Superior, and the Grenville, therefore, may be the time equivalent of the Keewatin iron formations farther west. The Hastings series they regard as Huronian.

Moore[3] describes the geology of the Sturgeon Lake goldfield in 1911. It is located in latitude 50° north longitude, northwest of Thunder Bay. The rocks are pre-Cambrian, largely igneous, including granites, gneisses, aplites, quartz-porphyries, rhyolites, hornblende-syenite, diorites, diabase, basalts, gabbros, and porphyrites and their metamorphic equivalents. There are small patches of arkose, graywacke, and dolomite. The relations of the igneous rocks to each other is partially known, but the age of the sediments and the igneous rocks and their relations are still uncertain. Free-milling gold veins are found at contacts, and in cleavage and fault planes and other openings. The gangue is principally

[1] W. G. Miller, "District of Patricia," *Ont. Bur. Mines*, XXI, Pt. 2, pp. 200, 18 figs., 29 maps.

[2] W. G. Miller and C. W. Knight, "Grenville-Hastings Unconformity," Abstract, *Geol. Soc. Am. Bull.*, XIX (1909), 539–40.

[3] E. S. Moore, "The Sturgeon Lake Gold Field," *Ont. Bur. Mines, 20th Ann. Rept.* pp. 133–57, 17 figs.

quartz, with small amounts of calcite and siderite. Sulphides of copper, lead, zinc, and iron are also present.

Moore[1] states that the rocks at Round Lake north of Lake Nipigon are Keewatin ellipsoidal greenstone, green schist, fine-grained quartose gneiss, and lean, magnetic, cherty iron formation, intruded by Laurentian granite and gneiss. The Archaean complex is intruded by Keweenawan diabase. The iron formation is separated from the greenstones by schists and sediments. It consists of bands of crystallized chert, magnetite, hematite, and some siderite.

Moore[2] states that the pre-Cambrian rocks of the Lake Savant Area northwest of Lake Nipigon belong to the Keewatin, Laurentian, Huronian, and possibly Keweenawan systems. The Keewatin comprises acid, intermediate, and basic extrusives and intrusives, banded magnetic, cherty iron formation, and an unusual development of sediments in the form of graywacke and gray gneiss. The Laurentian consists of granites and gneisses intrusive into the Keewatin. The rocks of the Huronian are principally conglomerates, graywacke, and a little quartzite. A few basic dikes may be of Keweenawan age.

Moore[3] describes the pyrite deposits of Big Vermilion Lake near the Minnesota boundary. The pre-Cambrian rocks of this region include a basement of intrusive and extrusive acid and basic rocks, overlaid by cherty iron formation, graphitic slate, graywacke, arkose and quartzite. The two systems were intruded by granite. The pyrite occurs in fissure veins in greenstone, quartz porphyry, and iron formation.

Moore[4] states that the Onaman Iron Range district covers about 70 square miles and lies between 45 and 50 miles up the Red Paint River, northeast of Lake Nipigon. The Keewatin, Laurentian,

[1] E. S. Moore, "Iron Range North of Round Lake," *Ont. Bur. of Mines, 18th Ann. Rept.*, XVIII, Pt. 1, pp. 154–62.

[2] E. S. Moore, "Lake Savant Iron Range," *Ont. Bur. Mines, 19th Ann. Rept.*, XIX (1910), Pt. 1, pp. 173–93, 16 figs.

[3] E. S. Moore, "Vermilion Lake Pyrite Deposits," *Ont. Bur. Mines, 20th Ann. Rept.*, 1911, pp. 199–208, figs. 6, 1 map.

[4] E. S. Moore, "Geology of Onaman Iron Ranges Area," *Ont. Bur. of Mines, Ann. Rept.*, XVIII (1909), Pt. 1, pp. 196–253, 22 figs., 3 pls., 1 map.

Huronian, Keweenawan, Pleistocene, and Recent rock systems are represented. The Keewatin consists of greenstone and greenstone schists, rhyolite, rhyolite porphyry, and feldspar porphyry, rhyolite tuff, agglomerate, and conglomerate, and iron formation. The Laurentian consists of granite and granite-gneiss, batholiths, intrusives into the Keewatin. The Keweenawan and Huronian are represented by intrusives. Pleistocene drift and Recent alluvial and travertine deposits cover a considerable portion of the district.

The Keewatin iron formation occurs in two nearly parallel, synclinorial belts running nearly east and west. Folding was accomplished by fracture and flow, predominantly the latter, although locally the formation is intensely broken and even faulted.

The various iron formation phases consist of cherty iron carbonate, ferruginous cherts, actinolite magnetite quartz rocks, interbedded with clastics whose grain suggests that they are of pyroclastic origin. The cherty iron carbonate and some of the iron oxide phases are probably original. The alterations of the iron formation have been chiefly anamorphic. Among the secondary minerals is dumortierite, a unique basic aluminium silicate, never before reported from iron formation.

Moore finds difficulty in accounting for the clean-cut banding of the iron formation as well as its sharp contact with the interbedded clastics on the hypothesis that the iron salts were contributed by the weathering of basic igneous rocks. These characters he finds in perfect accord with Leith's theory of direct igneous contribution. However, direct evidence for the latter he believes is lacking, since the iron formation is nowhere in direct contact with basic igneous rocks excepting at faults, and is also separated from rhyolite by slates, and what appear to be water-deposited pyroclastics, containing no iron formation. If the clastics in association with the iron formation are pyroclastic in origin, considerable difficulty in the way of the direct-contribution hypothesis would be removed. The iron and silica could have been delivered as well by weathering as by direct contribution in Moore's opinion. He cites the solubility of the silica in the iron formation as shown by Van Hise and Leith, the silica content of bog ores described by Hunt and himself as evidence of the adequacy of weathering in causing the deposition of silica.

While the evidence for the solubility of silica under weathering may be satisfactory, there is very little to show that weathering may yield siliceous deposits of anything like the thickness, high silica content, and freedom from detrital silica and clay as the Lake Superior sedimentary iron formations.

Moore believes that the Keewatin was a period of inclosed basins and imperfect drainage, conditions especially favorable to the solution, transportation, and deposition of iron. The long period of freedom from intense vulcanism, which he believes is implied by the thickness of the iron formation, removes the most apparent difficulty in assuming the development of a luxuriant flora, a condition conducive to the solution of iron and silica, and a selective retention of the complementary clastics on the land areas. Unless it can be shown that the clastics associated with the iron formation are pyroclastic in origin, it is obvious that a great deal of rock was disintegrated by weathering during the Keewatin. With due regard for the facts which may be interpreted as evidence for weathering and magmatic contribution, Moore concludes that the materials of the iron formation were supplied to inclosed basins by weathering action under the influence of plant life, and by heated igneous rocks coming in contact with the waters.

McInnes[1] reports that the pre-Cambrian rocks of the region between N. lat. 50° 10′ and N. lat. 55° 10′, and between W. long. 86° and W. long. 90°, include Keewatin diorites, diabases, and chloritic and hornblende schists, and coarse conglomerates, the latter probably Huronian, and Laurentian biotite granite gneisses.

Parsons[2] writes detailed descriptions of the gold prospects and mines of the Lake of the Woods, Manitou, and Dryden areas of northwestern Ontario.

Seelye[3] states that the Helen limonite ore body at Michipicoten is bounded by cherty iron carbonate, chert, and a diabase dike.

[1] W. McInnes, "Report on a Part of the Northwestern Territories Drained by the Umisk and Ottawapiskat Rivers," *Canada Geol. Survey*, 1910, 54 pp., 5 pls., 1 map.

[2] A. L. Parsons, "Gold Fields of Lake of the Woods, Manitou and Dryden," *Ont. Bur. Mines*, (1911) pp. 158–98, 26 figs., 4 pls.

[3] R. W. Seelye, "The Helen Mine, Michipicoten, Ontario," *Jour. Can. Min. Inst.*, XIII (1910), 121–34.

The ore body contains pockets of pyritic sand which have become more numerous with depth.

Wilson[1] reports on a reconnaissance survey of the area to the north of Lac Seul and east of Trout Lake in the Northwest Territories of Canada. He found only pre-Cambrian rocks, of which the oldest are chiefly hornblende schists and amphibolites. The greater portion of the area is underlaid by granites and gneisses intrusive into the basic rocks.

Wilson[2] finds that the pre-Cambrian rocks of the Algoma and Thunder Bay districts between 48° 30′ and 51° north latitude, and 84° and 87° 30′ west longitude, include Laurentian granite and gneiss with some hornblende schists, biotite schist, and diabase which may be Keewatin.

Wilson[3] states that the pre-Cambrian succession in the Nipigon river basin is as follows:

Keweenawan
- Diabase intrusives
- Dolomitic limestone
- Shale
- Sandstone
- Unconformity

Lower Huronian
- Small bands of conglomeratic arkose and slate
- Unconformity

Laurentian
- Granites and gneisses intrusive into Keewatin

Keewatin
- Greenstone and greenstone schists infolded with jaspilitic iron formation

Wilson[4] reports the following pre-Cambrian succession from the region about Larder Lake, Ontario and eastward:

Post Huronian
- Diabase, gabbro, porphyry, and lamprophyre
- Igneous contact

[1] A. W. G. Wilson, "Lac Seul to Cat Lake," *Canada Geol. Survey*, 1909, 23 pp.

[2] M. E. Wilson, "Geological Reconnaissance of a Portion of Algoma and Thunder Bay Districts, Ontario," *Canada Geol. Surv.*, 1909, 49 pls., 6 pls.

[3] A. W. G. Wilson, "Geology of the Nipigon Basin, Ont.," *Canada Geol. Surv. Mem. No. 1*, 1910, 152 pp., 16 pls., 4 figs., 1 map.

[4] M. E. Wilson, "Larder Lake and Eastward," *Canada Geol. Surv., Summary Rept.*, 1909, pp. 173–79, 1910.

Huronian
- Conglomerate
- Arkose
- Graywacke
- Conglomerate
- Unconformity

Laurentian
- Granite, gneiss, pegmatite, aplite
- Igneous contact

Keewatin
- Quartz-porphyry and porphyrite
- Rusty weathering carbonate rock
- Phyllite, slate, and graywacke
- Greenstone and green schist

Reviews

Climate and Evolution. By W. D. Matthew. Annals of the New York Academy of Science, Vol. XXIV, pp. 171–318.

This important paper is notable for the emphasis it lays on climatic variations and physical changes as agencies dominating organic evolution, for its adherence to the essential permanency of the continents, and for its unhesitating rejection of oceanic eversions and of extravagant bridge-building across abysmal depths for mere convenience in explaining biological distribution. In all these the author is loyal to the agencies attested by the geologic record and declines to go beyond them in summoning agencies of which the record gives no substantial authentication. He appeals to the powerful influence of climatic oscillations running back over the whole history of vertebrate life and beyond, whose verity is being constantly supported afresh by new evidence, and to the co-operative influence of physical changes connected with periodic diastrophism and denudation which have constantly varied the environment of life.

In addition to the long-recognized evidences of the essential permanency of the continents, the author cites the new support that springs from the geodetic evidences of isostasy and accepts all the difficulties this may seem to impose on the elucidation of biologic distribution. The author turns some of these supposed difficulties into evidences of the permanence of the ocean basins, by citing the obvious fact that a bridge between two great land masses should give a normal intermingling of the two faunas and floras inhabiting them and not a meager selection of forms susceptible of abnormal distribution by occasional modes of transportation. In this the author makes a notable contribution to the more critical study of what abnormal distribution really means. To the reviewer it seems probable that a really critical analysis of most cases of animal and plant distribution that are not in obvious harmony with the existing embossments and ridges, submerged or unsubmerged, will be found to imply selective transportation, not the normal commingling of species that naturally arises from a physical connection available to all species. The author's contribution to a closer scrutiny of the biological evidences of anomalous distribution is one of the first order of value. For its details the reader must consult the paper itself.

In his rejection of hypothetical eversions of ocean basins, the author does not hesitate to include Gondwana-Land, in some sense the parent—the mythical Atlantis aside—of the whole brood of lost continents. In this the author is likely to be sustained by stratigraphic and glacial evidence now already grown to a considerable mass which testifies, as well as bordering evidence well can, to the absence of any such elevated land connection between Australia, India, and South Africa across the Indian Ocean as is so commonly postulated to account for biologic distribution and the remarkable glaciation of those regions in Permo-Carboniferous times. In the opinion of the reviewer the stratigraphic and glacial evidences are not only quite against such intervening continent, but the hypothesis increases rather than alleviates the difficulties of explaining Permo-Carboniferous glaciation in those strange latitudes, and surely the difficulties are formidable enough without such hypothetical embarrassment.

The author does not push his negative attitude toward diastrophic extravagances farther back than Gondwana-Land because his study of vertebrate evolution does not seem to require it, but he might well have carried it back to the strange apparition of fishes and fishlike forms. Because coarse Devonian sediments with ichthyic faunas and terrestrial floras are found to encircle, in an imperfect way, the North Atlantic basin, an Atlantis is postulated in disregard of the strictly logical interpretation of both the physical and biological evidence which implies a disruption of the basin border, with a probable accentuation of the basin itself, instead of its eversion with the extremely improbable reversal of diastrophic action involved.

It is of course not improbable that some geomorphic changes of considerable importance have affected the ocean basins and the continental embossments in the course of geologic history, and possibly some of these may have been vital factors in the distribution of life, but the conservative example of the author in endeavoring to exhaust the probable influences of known fluctuations of milder type, attested by their own credentials, before resorting to collossal changes devoid of an appropriate record of their own, is wholesome and highly commendable. The paper should be read for its method as well as its material.

T. C. C.

The Noatak-Kobuk Region, Alaska. By PHILIP S. SMITH. Bull. U.S. Geol. Surv. No. 536, 1913. Pp. 157.

This report deals with the geology along the Noatak and Kobuk rivers which drain an area included approximately between 154° and 156° west longi-

tude, the Arctic Circle and 68° north latitude. The sedimentary rocks are divided into undifferentiated Paleozoic metamorphic schists, undifferentiated Paleozoic limestones, Silurian limestone, Upper Devonian limestone, Mississippian beds (Noatak sandstone, Lisburne limestone), a probable Jurassic Anaktuvuk group, a Lower Cretaceous or Upper Jurassic Koyukuk group, an Upper Cretaceous Bergman group, Eocene conglomerate, Pleistocene, and Recent. The igneous rocks consist of greenstones (pre-Carboniferous), probable Jurassic basic intrusives, flows, and tuffs, granitic intrusives (pre-Upper Cretaceous), and late Tertiary to Recent basaltic flows and volcanic ash. Deformation followed by erosion occurred previous to the deposition of the undifferentiated Paleozoic limestones. A period of deformation closed the Mississippian. The present topographic features were produced after the deposition of part of the Eocene. A series of east-west folds, broken by faults, affect the eastern part of the area, while north-south folds and faults are present in the western part. Auriferous gravels of notable production occur in the Shungnak and Squirrel River regions. There are gold and copper sulphide lodes, which have not as yet been developed to the producing stage.

V. O. T.

A Geological Reconnaissance of the Fairbanks Quadrangle, Alaska. By L. M. PRINDLE. With a Detailed Description of the Fairbanks District, by L. M. PRINDLE and F. J. KATZ. And an Account of Lode Mining near Fairbanks, by PHILIP S. SMITH. Bull. U.S. Geol. Surv. No. 525, 1913. Pp. 216.

The Fairbanks quadrangle, which forms a portion of the central plateau province of Alaska, lies between 64° and 66° north latitude and 146° and 150° west longitude. The most prominent structures are northeast-southwest trending close folds. The oldest rocks belong to the pre-Ordovician (?) Birch Creek schist; separated by an unconformity, a series of Paleozoic rocks (Ordovician to Carboniferous) follow. Unconformities also exist between the Silurian and early Devonian, between later Devonian and probable Pennsylvanian rocks, between the Upper Cretaceous and older rocks, and between the Upper Cretaceous and probable Eocene lignite-bearing beds. The occurrence of greenstones (basic volcanic lavas and tuffs), which are probably in the main of Devonian age, is interesting inasmuch as similar rocks appear in many parts of Alaska which are apparently of about the same age. The intrusion of granites and diorites succeeded the deposition of the Pennsylvanian strata. In one locality a basalt flow of probable Tertiary age was noted.

In the detailed description of the Fairbanks district are brought out the facts that this district lies wholly within the area of the Birch Creek schist, that great intrusions of diorite, granite, porphyries, and dikes occurred probably at the close of the Mesozoic, and that the auriferous and other deposits are genetically related to the intrusions. Gold placers have been developed on a

large scale in the Fairbanks district. These produced in 1912 gold to the value of $4,370,000 and silver to the value of $31,203. The development of the placers, according to F. J. Katz, has been chiefly by drift-mining methods, although open-cut mining is practiced in shallow ground. P. S. Smith in his discussion of lode mining notes that practically all of the developed veins are free-milling gold lodes, and that six properties are producing and have their own mills. The estimated production for 1912 from the lodes was $200,000. It is believed that this production will increase with the introduction of improved methods and more extensive development.

V. O. T.

The Koyukuk-Chandalar Region, Alaska. By A. G. MADDREN. Bull. U.S. Geol. Surv. No. 532, 1913. Pp. 116.

The Koyukuk-Chandalar region is bounded by 146° and 154° west longitude and the Arctic Circle and 68° north latitude. The main purpose of this report is to describe that portion of the area in which gold placers have been developed. The oldest strata are of sedimentary origin and Paleozoic age. These include the pre-Ordovician (?) Birch Creek schist, Devonian (?), and Carboniferous (?) beds. Cretaceous beds represent the Mesozoic sediments, and Tertiary and Quaternary deposits the Cenozoic. Granitic and dioritic intrusions (mostly of Mesozoic age) are associated with the Birch Creek schist; basic igneous rocks and tuffs occur with the Devonian (?); some volcanic rocks are present in the Cretaceous; the youngest igneous rocks are effusives, basaltic and andesitic lavas, and tuffs of Quaternary or Tertiary age. Unconformities are recognized between the pre-Ordovician (?) and Devonian (?); between the Carboniferous (?) and Cretaceous; between the Cretaceous and Tertiary; between the Tertiary sediments and the Quaternary or Tertiary igneous rocks; between these igneous rocks and the Pleistocene; and between the Pleistocene and Recent. The dominant structural lines trend a little north of east. The older schists are closely folded, while the Paleozoic series is thrown up into more open folds with many faults. The Mesozoic and Tertiary sediments are locally folded and faulted. Placer gold is, at present, the only mineral of commercial importance in the Koyukuk district. The gold occurs in the present stream deposits and bench deposits and has been derived chiefly from the Birch Creek schist. Surface mining methods are employed. In general, the yearly production of the Koyukuk district has gradually increased during the last ten years; the total estimated production up to 1912, inclusive, is $2,700,000. No lodes of commercial value have been found. In the Chandalar district probably "the most promise lies in its known quartz-lode gold deposits," since the placers are poor and local. The lodes are associated with diorite intrusions into the schist. Because of the difficulty of transportation of machinery and supplies, no producing mines are in operation.

V. O. T.